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**Abstract**

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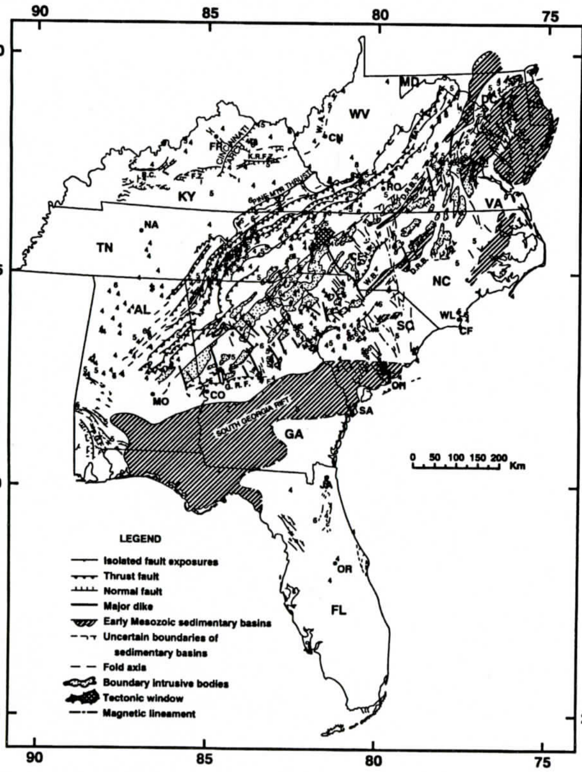
# SOUTHEASTERN GEOLOGY


GENERALIZED SEISMOTECTONIC MAP OF THE SOUTHEASTERN UNITED STATES





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# SEISMICITY AND STRESS PATTERN OF THE SOUTHEASTERN UNITED STATES

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## ABSTRACT

**An updated catalog containing 3050 earthquakes that occurred between 1698 and January 1995 is presented. The earthquakes appear to correlate with tectonics provinces, Cretaceous to Cenozoic faults, isolated fault exposures, boundaries of intrusive bodies, fold axes, faulted Triassic basins, linear magnetic anomalies, and dikes. A new seismotectonics map based on these data is proposed. The seismicity map shows several particular patterns:**

**1. an arcuate seismic zone in the northern parts of Kentucky, parallel with, but north of, the Rough Creek and Kentucky River fault zone;**

**2. the activity in the eastern Tennessee and northwestern Georgia related to the Valley and Ridge Province between the Pine Mountain Thrust Sheet and the Blue Ridge;**

**3. the Blue Ridge Province activity mostly around the perimeter of the Blue Ridge Uplift in eastern Tennessee and western North Carolina;**

**4. Piedmont Province activity in central Virginia associated with border faults of the Triassic basins and linear magnetic trends;**

**5. a major northwest-southeast seismic trend along the landward extension of the Blake Spur Fracture Zone in the Coastal Plain Province in South Carolina;**

**6. the epicentral position of the 1886 Charleston earthquake where an inferred volcanic series occurs, and where the northeastern border of the South Georgia Rift basin crosses the northwest-southeast seismic trend;**

**7. scattered seismic activities near or on the Pickens-Gilbertown Fault Zone, Goat Rock, Modoc, Gold Hill, Augusta, and Towaliga Fault zones;**

**8. minor seismic activity north of Orlando nearly parallel with, and near to, the eastern coast of Florida.**

**The spatial distribution of the events is interpreted in terms of seismic lineaments, or weak fault zones that are able to move sporadically under the regional stress field. Orientations of the zones are nearly northeast-southwest and northwest-southeast within the Piedmont and the Coastal Plain Provinces, but they are nearly north-south, and northeast-southwest within the Blue Ridge and Valley and Ridge Provinces. Azimuthal distribution of the zones shows a dominant north 30° east direction, while the azimuthal distribution of the available crustal stress orientations for this region indicates a dominant north 60° east direction. Assuming 30° for the rock frictional angle, the dominant orientations of the seismic lineaments appear to be consistent with the effects of the dominant direction of stress.**

## INTRODUCTION

The seismicity associated with the southeastern United States is diffused and rather infrequent. When causative structures are not recognizable, the concepts of seismic source zones and seismotectonics provinces are useful methods for seismic hazard estimation and risk evaluation. Seismic source zones in southeastern United states are discussed by Bollinger(1973), Rankin (1977), Gohn (1983), Talwani (1982, 1989), Hatcher and others (1977), Thenhaus and others (1987), Nuttli (1973, 1881), Hamilton(1981), Johnston and others(1987), Johnston (1987), Dewey(1985),

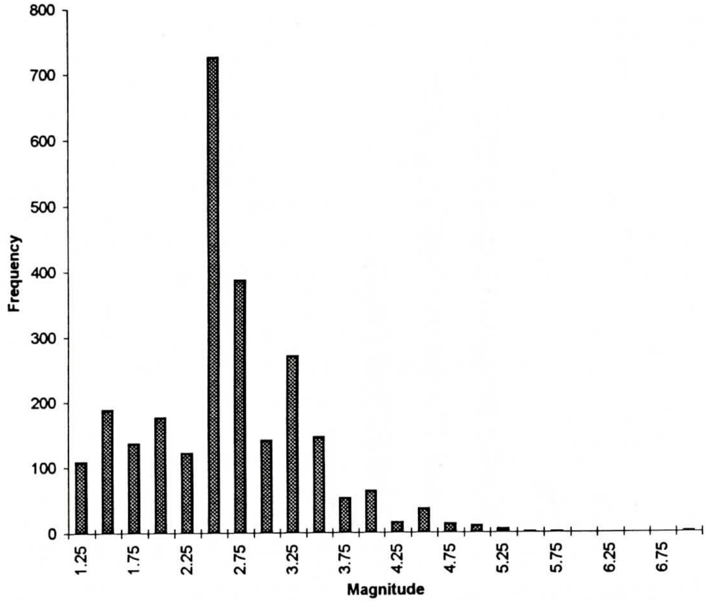


**Figure 2. Frequency distribution of $m_{bLg}$ magnitude for the southeastern United States. For the events prior to 1972 which did not have reported magnitudes but their intensities, I, were known, the statistical relation, $m_{bLg} = 0.656*I+0.402$, of Nowroozi(1991b) was used.**

the events are complete to $m_b = 4.5$ and for the period before 1930 the events are complete to $m_b = 5.7$ magnitude level. Using magnitude-intensity equation of Nowroozi (1991b) the corresponding intensities are III, VI, and VIII respectively. Thus, probably events with intensity IV are not complete for the period prior to 1930. In Figure 3 the intensities are written by the Arabic numerals instead of the usual Roman numerals for ease and clarity of presentation by computer graphics. The epicentral accuracy of historic events is not known. It is probably in tens of kilometers. The accuracy has improved since 1977, when the modern local seismic networks were deployed. The events located after 1977 probably have an epicentral accuracy of a few kilometers or less, and their focal depth may have less than 5 kilometers of error, (Bollinger and others, 1991). The area with mini-networks may have even higher accuracy, probably 1 km or less; but these accurately recorded events have very small magnitude or intensity; there are 2179 events with intensity less than III which are not considered for construction of the seismotectonics map. Although the epicentral positions are geometrical points, the symbols used to show them cover an area of about 50 square kilometers on the map. Thus, the error ellipses are not shown. A comparison of the seismicity maps based on historic data (1698-1977), and modern data (post 1977) show that they both have a similar spatial pattern, (Bollinger and others, 1991). But, temporal patterns indicate some distinction. The post 1977 data show a relative decrease in seismic activity in the northern Virginia Appalachian, and South Carolina Piedmont, but there are relative increases in activity in the northern Kentucky plateau and in the eastern Tennessee seismic zone, (Bollinger and others, 1991). This is the main reason that, I have used the entire data set from 1698 to 1995 in this paper. In my opinion, this data set (871 events) will be a better representation of both the spatial and the temporal pattern of regional seismicity than 293 events in the post 1977 data set. Various proposed geological source structures will be associated with the seismicity map, Figure 3, to find their seismic activity.

The relatively short duration of seismicity data, and nature of their accuracy have forced me to make an assumption: a geological struc-

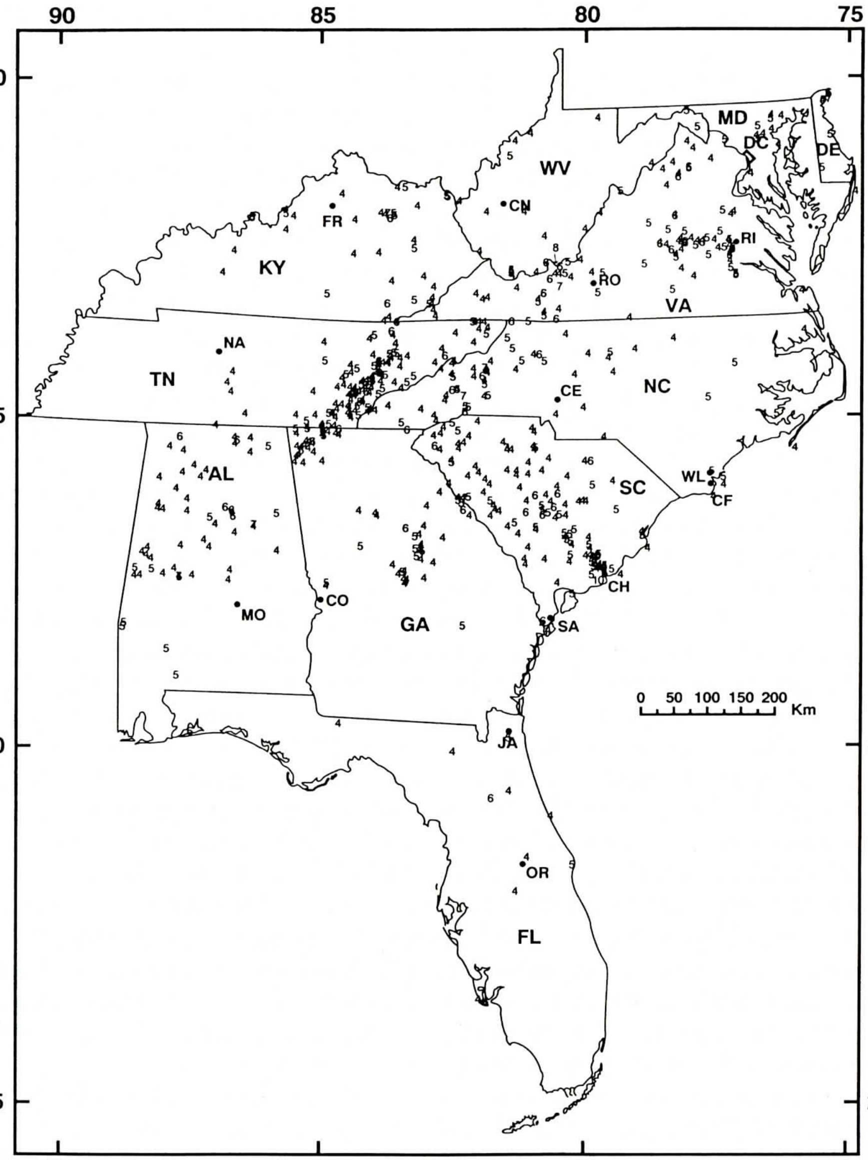


**Figure 3. The Seismicity map of the southeastern region of United States from 1698 to 1995. The epicentral positions of events are shown by Arabic numerals which indicate their seismic intensities in Modified Mercalli Scale. The name of states and several cities are shown by two letters abbreviations; the positions of cities are indicated by solid dots.**

ture is potentially seismogenic if some historic or instrumentally recorded epicenters are on it or are so close to it that, the structure may be suspected of having seismic activities. In the absence of Holocene faulting and lack of extensive and accurate historic and instrumental seismic data, I believe that, it is better to be at least moderately conservative and make this assumption, so that the potentially suspected seismogenic structures are identified. Owing to the scale of the maps, the spatial associations may be an approximations; however, this work delineated seismogenic structures which may require additional detailed work for confirmation.

## EARTHQUAKE CAUSES

It is commonly accepted that crustal earthquakes are produced when the state of stress around a volume of crustal block exceeds the crustal strength, and as a result the crustal block fails and faulting occur. Active Quaternary faults similar to those in the western United States have not been recognized or they are concealed in this area; thus, a number of other earthquake causes are proposed for the southeastern United States. Seismicity has been ascribed to granitic pluton (Kane, 1977; Sykes, 1978; Long and Chapman, 1977), reactivations of ancient deep fault (Behrendt and others, 1981; Seeber and Armbruster, 1988; Bollinger and others, 1991), reactivation of Mesozoic rift basins (Aggarwal and Sykes, 1978; Coruh and others, 1988), post-Cretaceous reverse faulting (York and Oliver, 1976; Wentworth and Merger-Keefer, 1983) northeast-southwest orientation of compression axis (Zoback and Zoback, 1980), intersection of major structural zones (Talwani, 1989a), neotectonic movement (Barosh, 1986, 1990; Talwani, 1991; Marple, and Talwani 1993), and concentration of stress near the boundary between relatively strong and the weak basement crustal blocks, Powell and others (1994). Talwani (1989a) presents criteria for zones of weaknesses in the crust. When there are weak zones, such as older faults, fracture zones, zones of intense dike activity, suture zone, terrane boundary, structures associated with the linear gravity or magnetic anomaly, they may be reactivated as the results of the regional stress field. Thus, under present regional stress patterns, the reactivations of crustal weak zones are the most likely cause of many recent low intensity events in this area. Therefore, the preliminary seismotectonics map is a compilation of faults, fault exposures, major dikes, boundary faults of the Mesozoic basins, fold axes, the boundary of large intrusive bodies, igneous dikes, and magnetic lineaments, as well as earthquakes with intensity higher than IV on Modified Mercalli scale from 1698 to January 1995.

## REGIONAL STRESS PATTERN

The state of stresses in the eastern region of the United States is discussed by a number of authors. Sbar and Sykes (1973, 1977) used several focal mechanism solutions and in situ stress measurements and deduced a northeast to east-trending maximum compressive axis direction. Bollinger and others (1991), and Talwani (1989b, 1991) presented the orientation of the P-axes from focal mechanism solutions of the 44 recent earthquakes. The averaged P-axes for 32 events in Eastern Tennessee, Giles County, Virginia, and Charleston, South Carolina is northeast-southwest. The complex seismic sources in Central Virginia show a mixed orientation. The averaged P-axes for four shallower events are northeast-southwest, and for seven deeper events are northwest-southeast, Bollinger and others (1991). The P-axis of one event in Kentucky shows east northeast orientation, (Herrmann and others, 1982). Zoback and Zoback (1980, 1981, 1989), and Zoback and others (1986, 1991) compiled the orientation and relative magnitudes of in situ tectonics stress in the continental United States from a number of stress indicators. Analysis of their data in this paper shows that the orientation of stress axes in the southeastern United State varies between north 40° and north 130° east, but there is a dominating north 60° east trend. This is in agreement with the N63°E direction obtained by Madabhushi and Talwani(1993).

## SEISMIC LINEAMENTS

In order to evaluate the effect of dominant direction of the stress orientation on seismic patterns in this region, I have interpreted the seismicity pattern in terms of many weak fault zones, or seismic lineaments. Barosh (1986, 1990), Sykes (1978), and Talwani(1989b) have suggested NW-SE and NE-SW trending alignments of epicenters in the southeastern United States. Barosh considered very long epicentral alignments along continuation of the Norfolk fracture zone in Virginia, along continuation of the Blake Spur fracture zone in South Carolina, and several other relatively long alignments with the same trend as faults or probable faults. If the lineaments are faults, they probably should have shown a major event in historic time. But, except for the Charleston event of 1886 and the Giles County events of 1897, there are no other large events in this area. Furthermore detailed seismic surveys (Ackermann, 1983) have failed to indicate the existence of northwest trending faults in the area of the Charleston event. I have interpreted the seismicity pattern in terms of smaller lineaments, which may be fractures produced in the crustal block in response to regional stress field. There are several linear alignments of epicenters with various orientations. The trends of the seismic lineaments vary from north 310° to north 40° east, with a dominating north 30° east direction. The New York-Alabama and the Clingman aeromagnetic lineaments which are associated with seismic activity in the eastern Tennessee seismic zones have about the same direction, Powell and other(1994). Based on this linear interpretation of the seismic pattern, and assuming a 30° angle for rock friction, the seismic activity appears to be consistent with an east northeast stress orientation. This trend is in agreement with the driving force of the ridge-push model for the motion of the North American plate, Reding (1984).

## GEOLOGICAL STRUCTURES AND SEISMICITY

A compilation of fault exposures, faults, border faults of the Triassic basins and major fracture zones, is presented in Figure 4. The geological structures will be correlated with the seismicity, Figure 3. The association of seismicity and faults is presented in Figure 5. A cross section of earthquake foci nearly perpendicular to the seismic trend within the Valley and Ridges and the Appalachian Mountain is presented in Figure 6.

The seismic data indicates that all the states within the southeastern region have sustained earthquakes with intensities higher than V; the

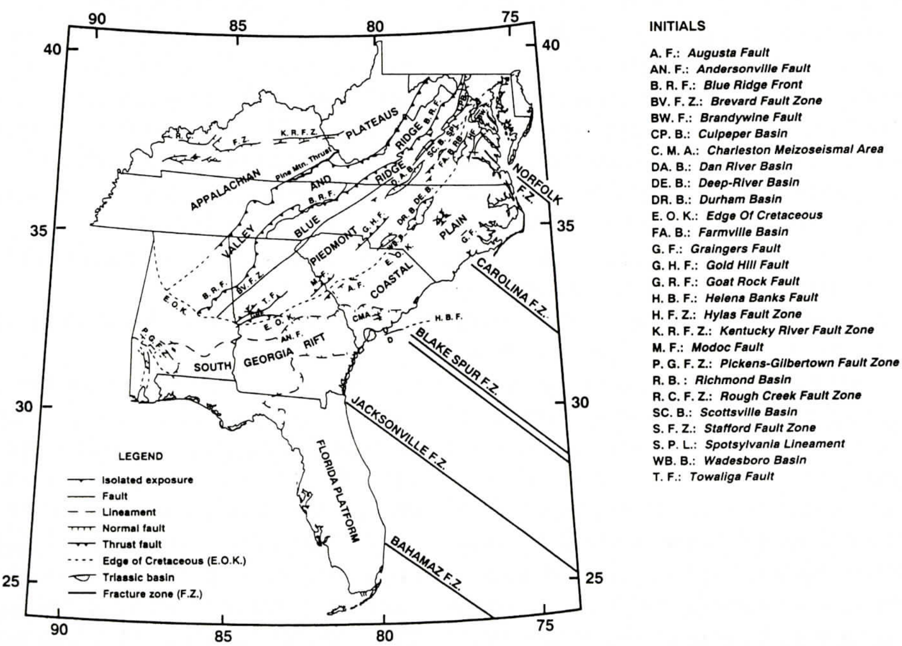


**Figure 4. Major faults, fault exposures, Triassic basins, fracture zones, and tectonics provinces in the southeastern region of United States. The sources are Cohee (1962), King (1969), Jacobeen (1972), Mixon and Newell (1976, 1977), Stover (1977), Howard and others (1978), Prowell and O'Connor (1978), Sykes (1978), Daniels and others (1983), Dischinger (1987), Hatcher and others (1977), and Prowell (1988).**

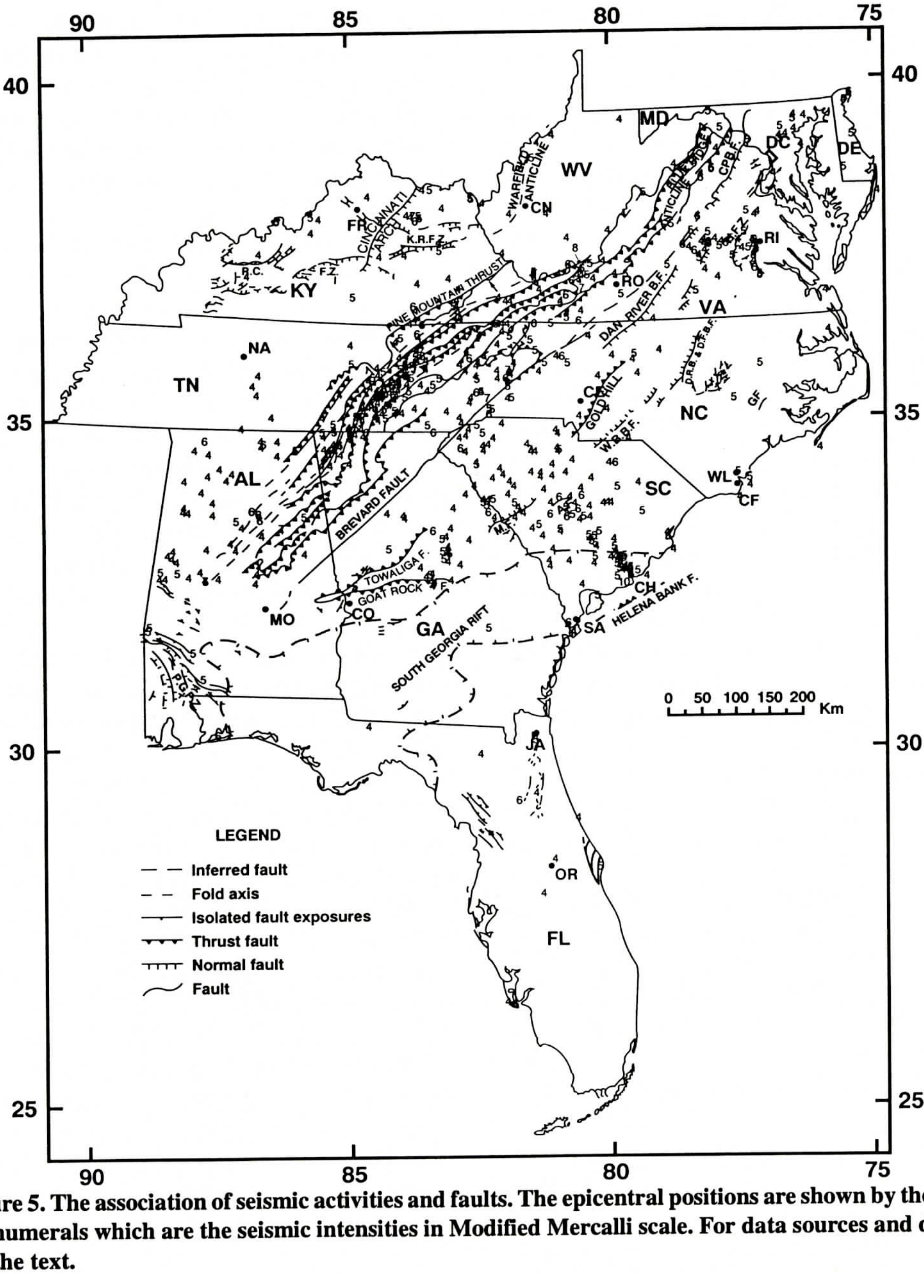


**Figure 5. The association of seismic activities and faults. The epicentral positions are shown by the Arabic numerals which are the seismic intensities in Modified Mercalli scale. For data sources and details see the text.**

maximum observed intensity is VI for the states of Florida, Georgia, West Virginia and Delaware, and VII for the states of Alabama, Kentucky and North Carolina. The maximum intensity of VIII and X are reported for the State of Virginia and South Carolina respectively. No earthquake with a maximum intensity of IX is reported for the entire area. The seismicity data shows several particular patterns; a summary is presented in this section.

**1.** There is a nearly arcuate seismic zone in the northern parts of Kentucky. The zone is nearly parallel with but to the north of Rough Creek and Kentucky river Fault zones, Figure 5.

**2.** In the eastern Tennessee and northwestern Georgia, the majority of events are between the

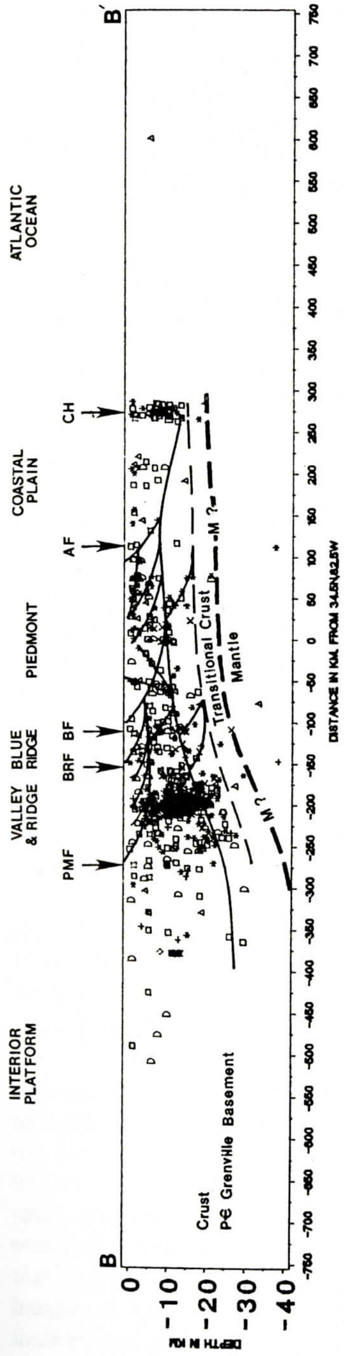


**Figure 6. The cross section of earthquake foci nearly perpendicular to the seismic trend within the Valley and Ridges and the Appalachian Mountain. The horizontal distances are measured from a point in the state of South Carolina with coordinate of: Latitude = 34.5° N; Latitude = 82.5° W, the trend of profile is N 45° W. All the events within 200 Km. width which have known focal depth and intensities higher than I are plotted. The positions of Pine mountain thrust, PMF, the Blue Ridge Front, BRF, the Brevard fault, BF, the Augusta fault, AF, and the Charleston seismic zone, CH are given. The heavy lines are possible fault traces drawn through the alignment of the focal positions. The deep events, between PMF and BF are below the indicated thrust faults, but they are within the Precambrian Grenville crustal rock.**

Pine Mountain thrust sheet and the Blue Ridge. The earthquakes are related to the Valley and Ridges Province, Figure 4 and 5. Their foci are relatively deeper than the foci in other areas, Figure 6. Johnston and others(1985) also have suggested that the Tennessee seismicity is associated with the basement crustal structures at depth of about 7 to 25 km. Powell and others(1994) suggested that the cause of seismicity is the concentration of the regional stress field near the boundary between relatively the strong and the weak basement crustal blocks.

**3.** Between the Brevard fault zone and the Blue Ridge front, the Blue Ridge province shows moderate seismicity, Figure 4 and 5.

**4.** Within the Piedmont Province in central Virginia, Figure 4, the majority of seismic activity is probably associated with the border faults of the Richmond and the Farmville Triassic basins, as well as dikes and magnetic lineaments, Figure 5, 7 and 8.

**5.** Within the Coastal Plain province in South Carolina, there is a major northwest-southeast seismic trend along the landward extension of the Blake Spur Fracture Zone, Figure 4 and 5.

**6.** The northeastern border of the South Georgia Rift Basin crosses the northwest southeast trending seismic zone nearly at the epicentral position of the 1886 Charleston earthquake, where the inferred buried volcanic series (Ragland and Hatcher,1983; de Boer and others 1988) are also reported, Figure 5 and 7.

**7.** There is scattered seismic activity near or on the Pickens- Gilbertown Fault Zone, Goat Rock, Towaliga, Modoc, Augusta, and Gold Hill Faults, Figure 5.

**8.** There is minor seismic activity north of the Orlando township parallel to and near the coast of Florida, Figure 5.

## SEISMICITY OF THE MESOZOIC AND CENOZOIC FAULTS

The Quaternary faults are the foremost candidates for seismic sources, however so far they are not recognized in this region, (Wentworth and Mergner-keefer, 1983). Based on geological mapping, interpretation of recent aeromagnetic maps, and field checking, Hatcher and

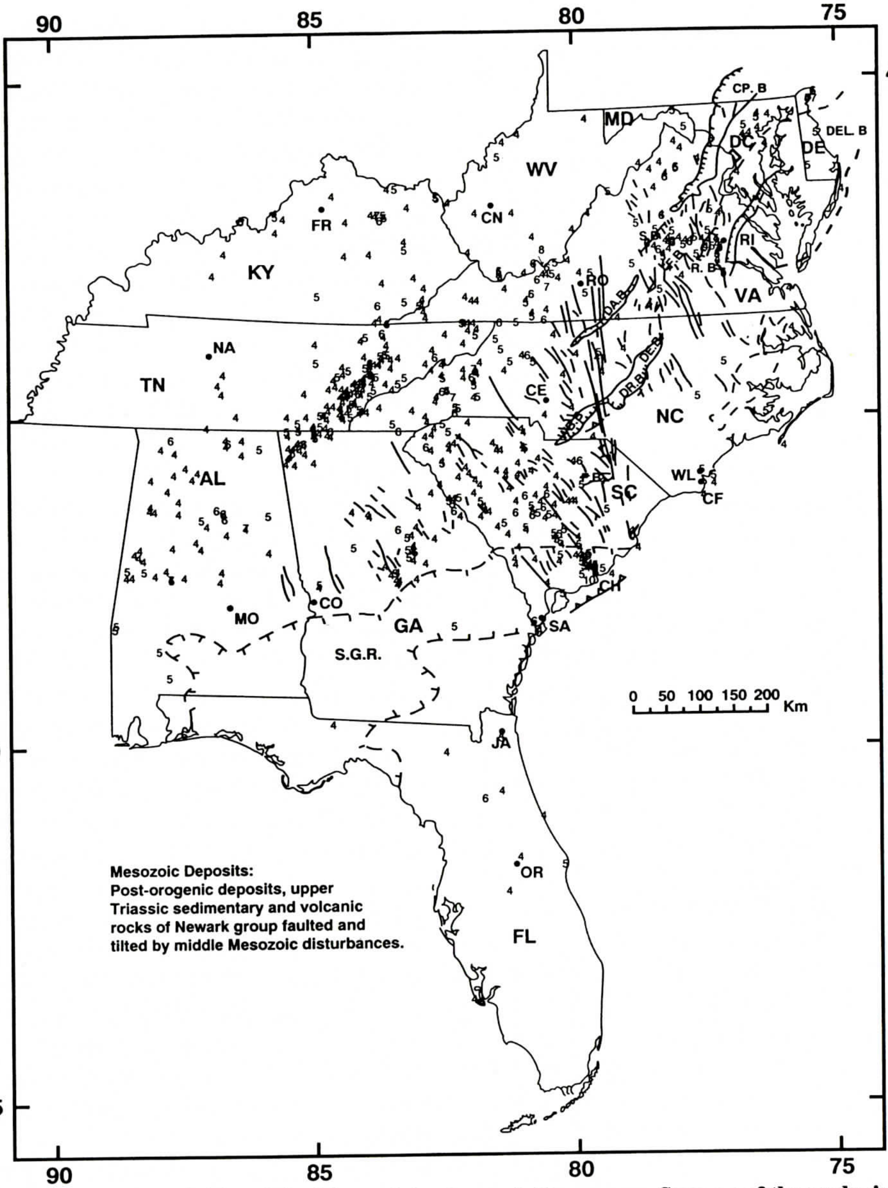


**Figure 7. The seismic activities of the Mesozoic basins and dike swarm. Sources of the geological data are de Boer and others (1967), Wentworth and Merger-Keefer(1983) Daniel and others (1983). For details see the text.**

others (1977) speculated on the extent of the eastern Piedmont fault system. They defined this system as related series of linear magnetic anomalies associated with cataclastic zones of varying thickness and dip which also included several known faults. The system includes the Goat Rock fault in Alabama and Georgia, the Modoc, and Augusta fault in South Carolina, border fault of the Durham basin in North Carolina, and the Hylas fault in Virginia. They also noted seismicity associated with parts of this system. We will discuss seismicity associated with several segments of the Eastern Piedmont fault system later. The other major fault system in this area is the Brevard fault zone, (Bobyarchick and others, 1988). The fault zones extend association with the Rough Creek, Kentucky River fault zones, the Cincinnati arch and the Warfield anticline, Stover (1977).

Herrmann and others (1982) reported a right-lateral strike slip motion on a north 30° east nodal plane for the Sharpsburg, Kentucky event of 27 July 1980, Latitude = 38.17° N, Longitude = 83.91° W, magnitude = 5.2, and intensity = VII. The compressional axis for this solution is northeast southwest; however for the two aftershocks solutions, one mechanism gave a similar compressional axis, but the other solution gave an east-west compressional axis. A large number of events with intensity of IV to VIII occur between the Pine Mountain thrust and the Brevard fault zone in the Valley and Ridge and the Blue Ridge tectonics provinces. The intensity VIII belongs to the Giles County events of 31 March 1897, Latitude 37.03° N, Longitude = 80.70° W, magnitude = 5.8. Bollinger and Wheeler (1982) and Bollinger (1981) reported that the recent seismic activity in the meizoseismal area of this event have occurred on the steeply dipping buried fault zone oriented north northeast. Furthermore they inferred that a late Precambrian Cambrian normal fault formed during the opening of the Iapetan Ocean and its reactivation by the present stress field may be the likely seismogenic source.

The faults, within the border of the Pine Mountain thrust and the Brevard fault zone, are interpreted as shallow angle thrusts with relatively small depth by Harris and Bayer (1979) based of seismic reflection data. From the records of a dense seismographic network, Johnston and others (1985) found that most of the seismic activity in this area occurs below a decollement and are unrelated to the surface geology. Furthermore, they determined several high quality focal mechanism solutions that indicates a right lateral strike slip motion on a nearly north south striking nodal plane and a northeast-southwest orientation for the compressional axis. Powell and others (1994) presented a model for the eastern Tennessee seismic zone, they suggested that most of the events occurred in the crystalline basement rocks in response to the intraplate stresses that are now concentrating near the boundary between a weak and a strong crustal blocks. Data presented in this paper also shows that the seismic events in this region are confined to a depth of 12 to 25 km, well below the thrust planes. Thus, correlation of the thrust faults with the seismicity is not established in this area for the deeper events.

For more recent events, which are recorded instrumentally, focal depths are available. A profile of focal positions for all events with higher than intensity I within a 200-km zone perpendicular to the Appalachian seismic trend is given in Figure 6. The horizontal distance is measured from a point with coordinate of 34.5° north, 82.5° west. The vertical axis is five times larger than the horizontal axis; thus, dip angles are exaggerated. The positions of the Pine mountain thrust, the Blue Ridge Front, the Brevard fault, the Augusta fault, and the Charleston seismic zone, as well as my interpretation of fault traces are shown in the Figure 6. This interpretation is similar to that given by Johnston and others (1985). Several events with intensity of IV to VII occur on the northeastern segment of the Brevard fault zone and the Grandfather tectonics window in western part of the North Carolina; however, there is no seismicity associated with the southwestern segment of the Brevard fault zone so far. The Brevard fault zone is an important marker of the seismic regime in the southeastern seaboard. To the southeastern part of it, the seismicity is dispersed and earthquakes have relatively shallower depth, but to the northwest part of it, between this fault and the Pine Mountain thrust, the seismicity is more concentrated and earthquakes have greater depth, Figure 6.

Several isolated fault exposures with Cenozoic offsets have been reported in the Coastal Plain area of the southeastern United States by Prowell(1983), Mixon and Newell (1976), and Prowell (1988). There are minor seismic activities associated with some of them. Some events with intensity IV and V are located on or in close proximity of these fault exposures in northern Virginia, near the Spotsylvania lineament, and the Stafford fault zone. There are several intensities IV to VI events on a north-south trending fault exposure south of Richmond.

Several events occur on the fault exposures in the meizoseismal area of the Charleston earthquake of 1886. In addition, two events occur on an exposure in the Jacksonville area in northern Florida.

## SEISMICITY OF THE MESOZOIC BASINS AND DIKES

Continental rifting and crustal thinning took place between the North American and African continent, during Late Triassic and Early Jurassic. Numerous early Mesozoic continental basins were formed before and during rifting when crustal segments were pulled apart. These basins, often known as Triassic rift-basins, developed into pull apart basins, half grabens or rhombi-shaped full grabens as transform segments developed during episodes of crustal extension and lateral compression, Manspeizer (1981). The border faults of these basins are preexisting weak zones of the crust which may be seismogenic under the present compressional stress system. The Mesozoic border faults are suspected of the neotectonics reactivation, their seismic activities are previously reported, (Aggarwal and Sykes, 1978; Coruh and others, 1988). The locations of the exposed and the approximate limits of the buried Triassic basins, and dike swarms, de Boer and others (1988), together with the seismicity are given in Figure 7. Sources of data for the basins are, Wentworth and Mergner-Keefer (1983), Higgins and Zietz(1983) and Daniel and others (1983). The border faults of exposed Culpeper, Richmond, and Scottsville basins appear to be associated with earthquakes of intensity V or less. The border faults of the unnamed buried basins in the state of Delaware, Virginia, and eastern part of North Carolina, may also be associated with intensity V or less events. There are other events with similar intensities in close vicinity of the Dan River, Farmville, Durham and Wadesboro basins.

Based on potential data, and core samples Daniels and others (1983) have defined the border of the south Georgia rift basin which extends from the Gulf Coast of northern Florida to Georgia and southern part of South Carolina. The Helena Banks fault, (Behrendt and others, 1983), and the meizoseismal area of the Charleston earthquake of 1886, Wentworth and Mergner-Keefer (1983), are located in the northeastern part of this basin where a volcanic series was inferred from the magnetic signature and several core samples, de Boer and others (1988). The epicentral position of the Charleston earthquake of September 1, 1886, Latitude = 32.9° N, Longitude = 80° E, magnitude = 6.9, moment magnitude = 7.3, Johnston, 1995, and intensity = X, is near the intersection of the northwest trending seismic zone in the state of South Carolina and the northeastern border of the South Georgia Triassic rift basin. Talwani (1982) found that the relocated events and the composite focal mechanism solutions in the meizoseismal area of this event delineate two main seismic zones lying at different depths. The shallower zone showed a reverse faulting on a steeply dipping northwest nodal plane, while the deeper zone showed a strike slip faulting on a northeast trending nodal plane; both mechanism solutions were consistent with a north 60° east orientation for the compressional axis. Recent work of Madabhushi and Talwani (1993) indicates a N63°E direction for stress orientation. Furthermore, they postulated that the concentrated events are occurring at the intersection of the Ashley River and the Woodstock fault zones. Within the top 750 m, several small faults, believed to be part of the border fault of the basin, are identified by the seismic refraction method in the meizoseismal area of this event, Ackermann (1983). The remaining part of the basin or its border has only sustained events with intensity of IV to VI in the states of Georgia, Alabama and Florida.

The positions of the inferred dikes from magnetic lineaments, Daniels and others (1983), and de Boer and others (1988), with the seismicity data are also given in Figure 7. From a combination of geophysical and borehole data, de Boer and others (1988) also inferred the Charleston volcanic series of the South Carolina, Figure 7. This series is observed as a reflector with high velocity, (Dillon and McGinnis, 1983). The meizoseismal area of the Charleston earthquake of 1886 is well within the boundaries of this volcanic series. A large number of dikes are also inferred based on their magnetic signatures on lands. They radiate outward from the Charleston volcanic series, Figure 7. The dominant orientation of this series varies from northwest in the south, to north-south in the central and to northeast in the north Appalachians, King (1961, 1971), and de Boer (1967). The emplacements of the dikes are believed to be related to fractures, mostly reactivated faults as a result of regional left lateral shears, in the late Jurassic, de Boer (1967), or counter clockwise rotation of North America with respect to Africa in middle Jurassic, Swanson (1982).

In the state of Georgia a group of intensity IV events appears at the intersecting terminus of two dikes, and one intensity IV event is on another dike. But the remaining majority of the events are not on any dikes. In state of South Carolina, dikes have a dominantly northwest and north-south orientations in line with the trend of a broadly observed seismicity pattern in that region and the extension of the Blake Spur fracture zone; there are several events with intensity IV to VII on or near terminuses of the dikes. In the state of North Carolina, the longest dikes have a north- south trend, there are a few events of intensity IV to V near their terminus or on them. In central part of Virginia, where the Central Virginia seismic zone occurs, (Bollinger, 1973), the trends of dikes show more complexity. Thereby, the seismicity data is plotted on a detail aeromagnetic anomaly map for this area, Zietz and others (1977), Figure 8. Association of linear magnetic anomaly with earthquakes is reported in the eastern Tennessee seismic zone, Powell and other(1994). They associated the New York-Alabama and the Clingman magnetic lineaments in the Valley and Ridge and Blue Ridge provinces with earthquakes. There are several other linear magnetic features, probably dikes or faults which are seismogenic. Two north-south lineaments are marked by several events in the east of Richmond; their intensities are up to VI. There are events clearly related to the Hylas border fault of the Richmond Triassic basin with intensity up to III; and three events are within the basin. Several events mark another north-south lineament west of the Sabat amphibolite complex which has a nearly north-south elliptical magnetic signature west of the Richmond basin. In addition, there are northwest and northeast trending magnetic lineaments with events of intensity up to VII on them or near their terminus. Coruh and others (1988) reported on the correlation between the earthquake foci in the central Virginia seismic zone with several reflectors seen on the reprocessed I-64 seismic-reflection data obtained in Virginia. They interpreted the lack of a continuous reflection at depth as dike swarm and mafic rocks. The top of this dike swarm was indicated as an antiform by a strong reflection which was dipping toward the west and the east outside of the dike swarm's limits. They reported that the shallower foci showed excellent correlation with the west dipping antiform and may be related to the reactivation of the thrust defining the roof or the faults above the structure. But the foci of the deeper events appeared to be related to the near vertical diabase dike swarm of the Mesozoic age. Their focal mechanism solutions for the shallower antiform events showed a northeast southwest trending compressional axes, but for the deeper dike events the axes were in the northwest south east direction. Thus, at least in the central part of Virginia several seismogenic magnetic lineaments appear to exist, Figure 8. The lineaments may be seismogenic faults or dikes.

## SEISMICITY OF INTRUSIVE BODIES

Sykes (1978) concluded that the seismicity of the southeastern region may be explained by preexisting zones of weakness, where alkalic intrusive bodies can also be found. Long and Chapman (1977), and Campbell (1978) considered the association of plutons and earthquakes in this region as results of stress amplification due to differences in the stiffness of plutonic inclusions and of the elastic crust. Kane (1977) explained the correlation between several large seismic events, including the Charleston event, and the anomalous mafic/ultramafic rocks as results of serpentinization. He argued that under stress, the serpentinized gabbro and dunite fail in creep mode rather than in strike-slip mode.

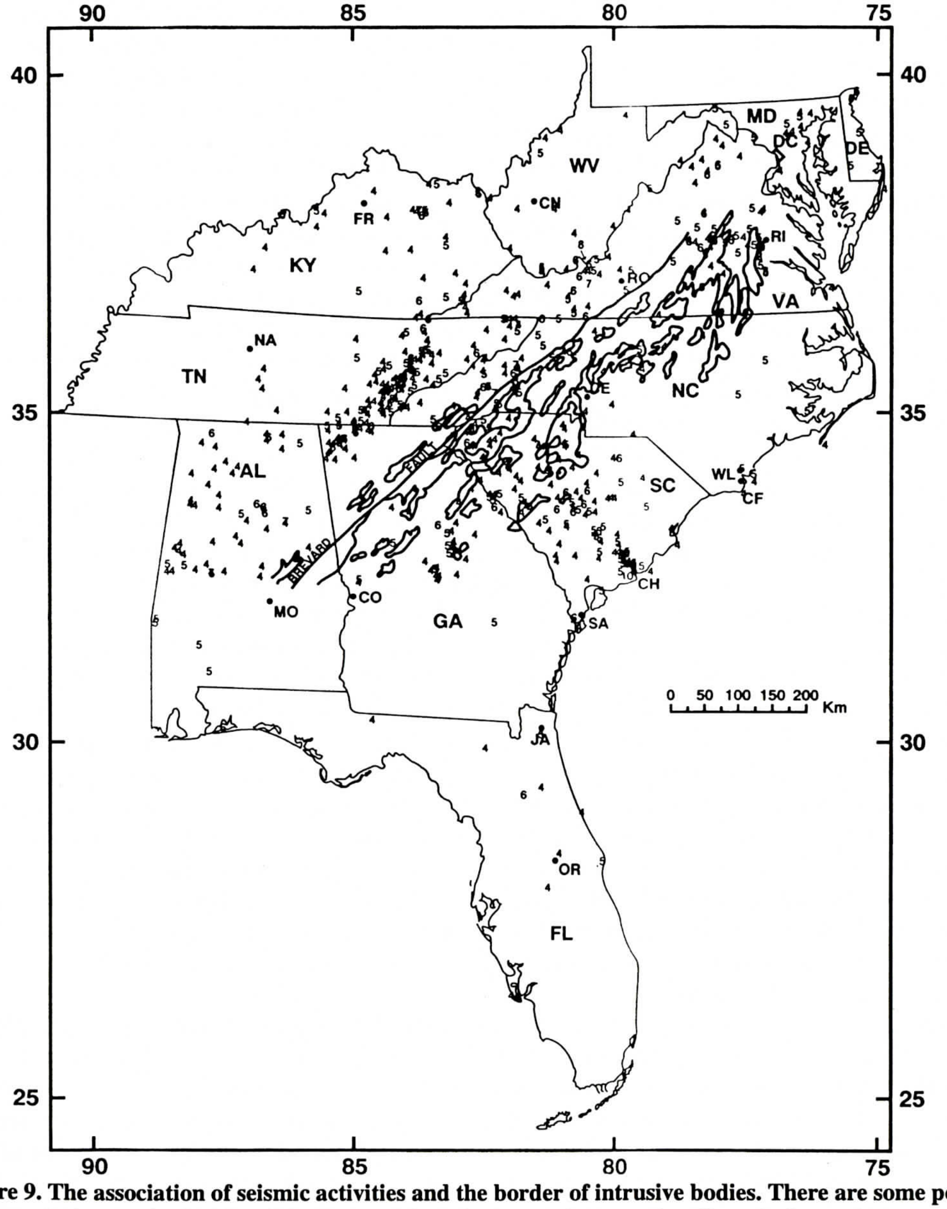


**Figure 9. The association of seismic activities and the border of intrusive bodies. There are some positive correlations on both sides of the Brevard fault in the northeast region Georgia, in northern area of South Carolina and the western part of North Carolina, and in central Virginia. The meizoseismal area of the 1886 Charleston earthquake is close to the border of an inferred volcanic series, de Boer and others (1988).**

Thus, the serpentinized rock can sustain low-stress field, therefore acting to concentrate regional stress outside their boundaries. McKeown (1982), however, considered this association the results of crustal weakening in the process of magmatic intrusion. Therefore, the boundaries of intrusive bodies may have some associations with the seismicity pattern. In Figure 9, the boundaries of intrusive, granitic and mafic plutonic rocks, Cohee (1962), and King (1969), and the post rift volcanic series in vicinity of the Charleston, de Boer and others

(1988) are correlated with the seismicity. Positive visual correlations are seen in the Piedmont Province of Virginia, in the western North Carolina and the northwestern part of South Carolina near the southeastern side of the Brevard Fault zone. Nowroozi (1991b), and Wheeler and Bollinger (1984) made an attempt to correlate seismicity patterns with the accreted terranes, Williams and Hatcher (1983), and Horton and others (1989). They reported some correlation between the broad terrane boundaries and seismicity pattern. Nowroozi (1991a), used gravity gradient, magnetic signature, and seismicity, in addition to other geological data and defined the Charleston terrane which includes the meizoseismal area of the Charleston earthquake. This area is within the post rift volcanic series of de Boer and others (1988).

## SEISMOTECTONICS MAP

Some elemental features and definition of the seismotectonics map are discussed in the literature. Hadley and Devine (1979) stated the purpose of a seismotectonics map is the presentation of historic seismicity in terms of geological structures and tectonics provinces characterized by a consistent relationship between seismic activity and structural features. Talwani(1989b) has defined the seismotectonics as integration of seismicity, geological and geophysical data with a view of defining and understanding the nature of seismogenic features. Nowroozi (1991a) has defined a seismotectonics province as a region characterized by a relative consistency of geological structural features, geophysical pattern and potential seismogenic sources contained therein.

In the previous part of this work I discussed the visual correlation of earthquakes with several classes of structure: faults, fold axes, isolated fault exposures of the Cretaceous-Tertiary age, boundaries of plutonic rocks, dike swarms of middle Jurassic, border faults of the Triassic age, and magnetic lineaments. I observed and reported on some positive correlations between seismicity and some members of the class of geological structure or their geophysical signatures. Although there has not been a positive correlations among all members of each class and seismicity, I believe that, it is prudent to assume the entire class as seismogenic, even if a few members of that class show some degrees of association with the observed seismicity, because the observed duration of historic and instrumental seismicity is only about 297 years. Thus, in a longer time duration seismic events may show on all members of that class. Thereby, I have presented some generalized and preliminary seismotectonics map for the southeastern region of the United States based on my interpretation of these data and the above assumption. The map contains seismic epicenters shown by their intensity in Arabic numerals, isolated fault exposures, faults, fold axes, major geologically mapped dikes, magnetic lineaments which are interpreted as dikes, and boundaries of the Mesozoic sedimentary basins, tectonics windows, and the intrusive bodies. Although the majority of seismic events are related to the structures discussed before, as can be seen from Figure 10, still there are a few events which are unexplained. For example, three events north of Cape Fear and few events in the southwestern area of Alabama are not near any structural features. But three events with intensity IV in central Tennessee are probably related to the Nashville dome, and two events with intensity V in the state of North Carolina are off a buried basin.

## ORIENTATION OF TECTONICS STRESS FIELD AND SEISMICITY PATTERN

The stress map for the southeastern area of United States, Figure 11, is redrawn from the stress map of North America by Zoback and others (1991). A frequency distribution of the data for the southeastern region indicates, that a majority of the stress orientations have a variation between north 40° east and north 130° east; however, a dominant direction of north 60° east is clearly observed, Figure 12. This direction is in good agreement with the northeast direction of the P-axes of recent focal mechanism solutions, Bollinger and others (1991), and N63°E direction reported by Madabhushi and Talwani(1993).

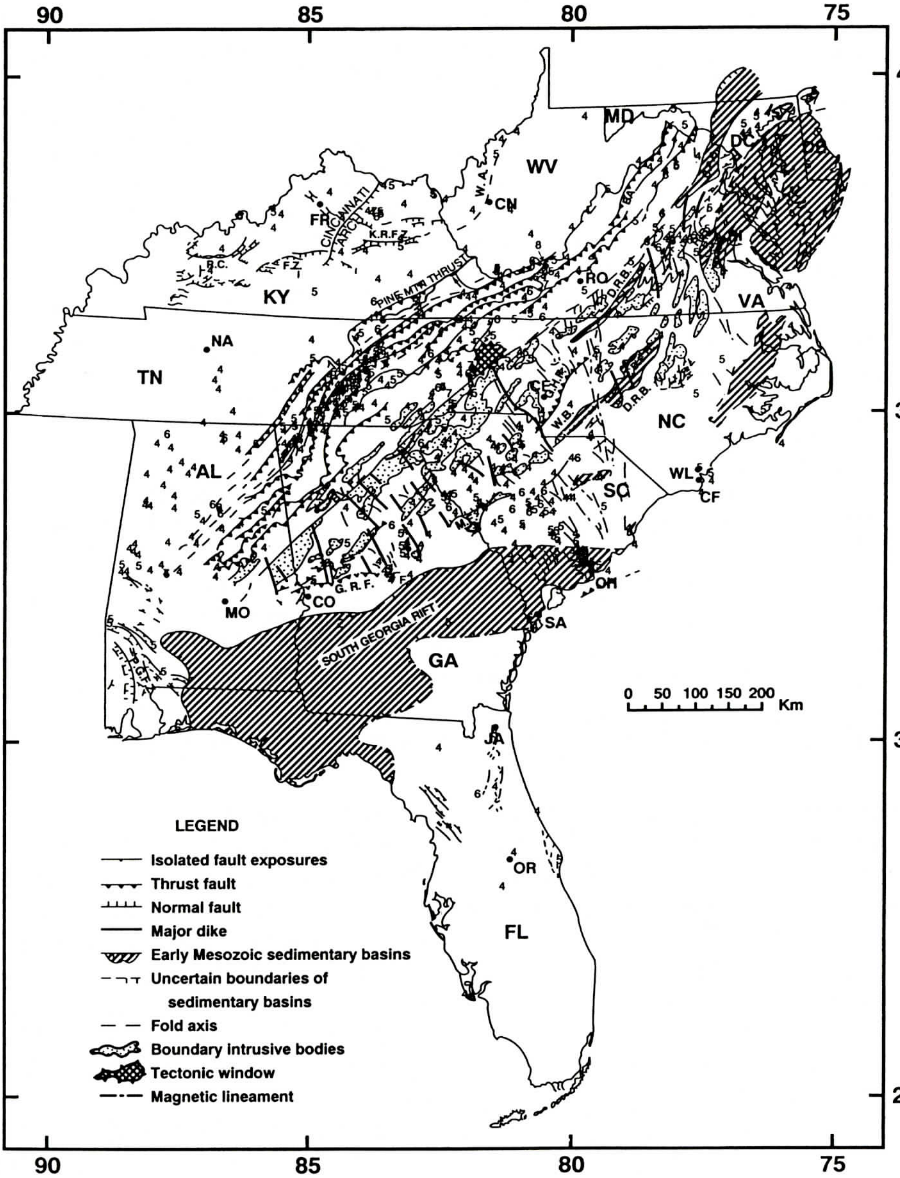


**Figure 10. The preliminary seismotectonic map of the southeastern region of the United States. Note although, a great majority of the events are on or in close vicinity of a geological structure or magnetic lineament, still there are a number of events which are not on any known structure.**

Barosh (1986, 1990) stated that most of the seismic sources in the eastern part of the United States appear to be related to northwest-trending fracture zones that commonly have apparent right-lateral strike slip offset, where they cross a broad northeast-trending belts of vertical movements. In addition, he reported on earthquakes which are related to north-trending extensional fault zones. Talwani (1989c) postulated nucleation of the intraplate earthquakes at intersections of preexisting zones of weakness as a working hypothesis. He also considered landward extension of the northwest-trending Blake Spur, and Norfolk fracture zone, Figure (4), with the northeast trending fault systems as areas with high seismicity. In

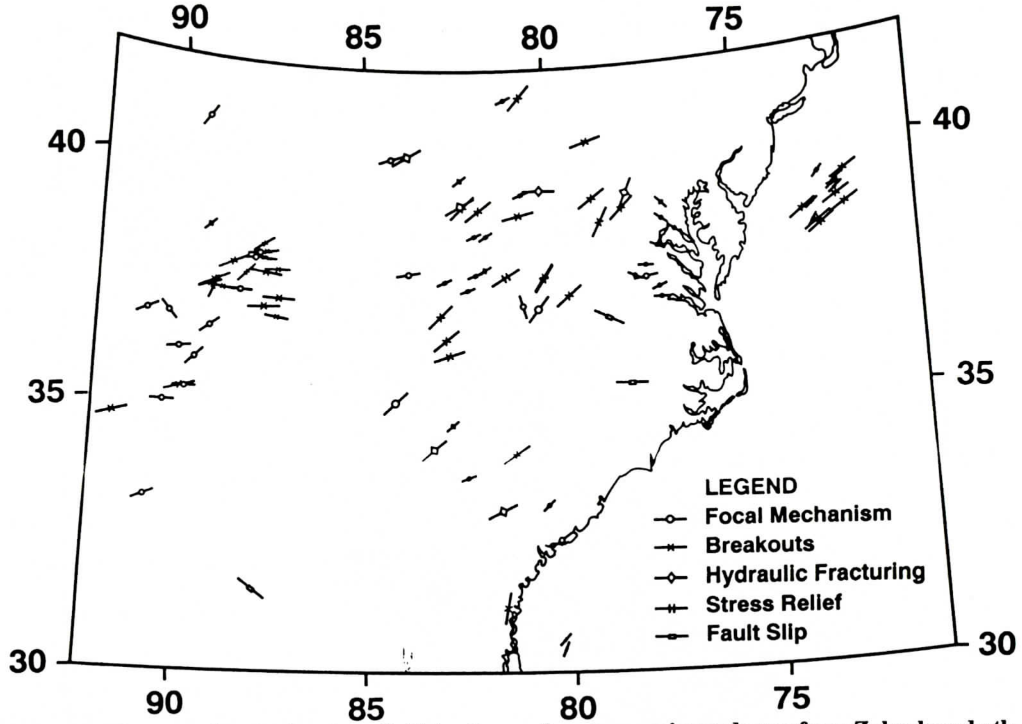


**Figure 11. The map of tectonics stress field for the southeastern region redrawn from Zoback and others (1991).**

the previous part of this paper, the north-south, northeast-southwest, and northwest- southeast trends of weak zones and seismicity were discussed. McKenzie (1969) argued that, the stresses causing shallow earthquakes and the occurrence of earthquakes along fault plane suggest that events occur by failure on weak planes, rather than by fracture of a homogeneous crustal material. Expanding on these ideas, I have evaluated the relation between the seismic activity and stress system by interpreting the seismic pattern in terms of short linear zones which are marked by seismic events, Figure 13.

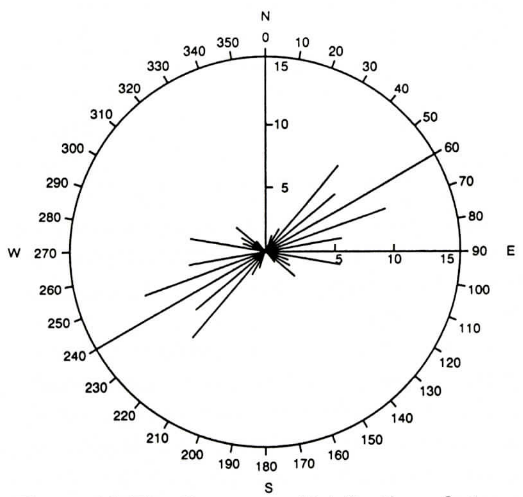


**Figure 12. The frequency distribution of stress orientation based on data in Figure 11. Note the prominent N60°E orientation of the stress field.**

About seventy short line segments of various direction are drawn through the epicentral position of earthquakes, these lines or seismic lineaments, are considered to be the intersection of the weak planar zones within the earth's surface. As Talwani(1989c) postulated, a high concentration of events appear to occur where the weak zones or the seismic lineaments intersect, Figure 13.

The frequency distribution of the seismic lineaments is given in Figure 14. A majority of directions are within the north 310° east to north 30° east, with a prominent trend at the north 30° east. The trend of the east Tennessee seismic

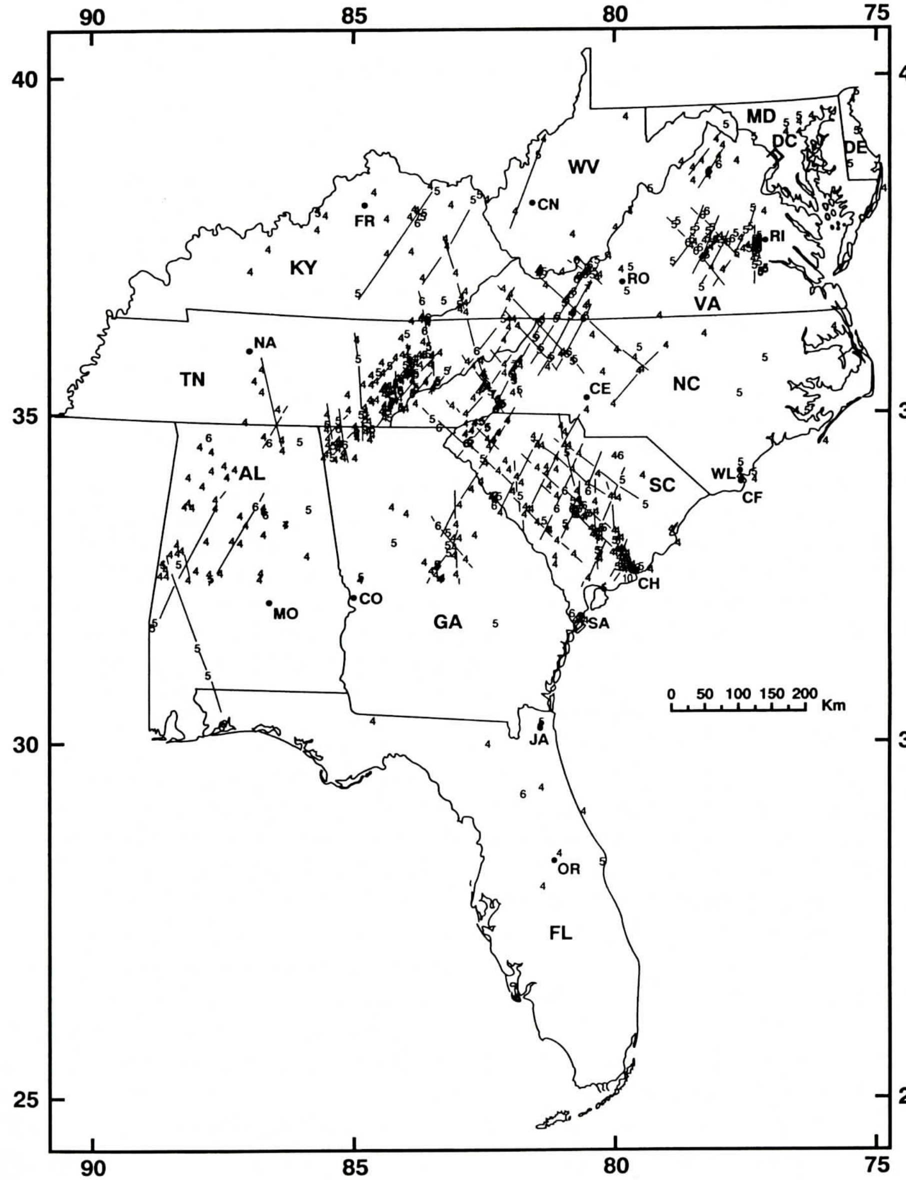


**Figure 13. The seismic lineaments map. The seismic activities are interpreted in terms of linear fractures, or weak zones, marked by epicenters. About seventy short lines with various orientations are drawn through the epicentral alignments. The linear fractures are called seismic lineaments. Note the concentration of activities where the lineaments cross each other.**

zone is also about N30°E, Powell and others(1994). In accordance with the dynamic theory of faulting, Anderson (1951), Hubbert (1951) and Hafner (1951), the maximum shearing stress occurs at $\beta = \pm(45+\phi/2)$, where $\beta$ is the angle from direction of minimum principal stress, and $\phi$ is the internal angle of rock friction. Assuming an average $\phi = 30°$, the conjugate shears may develop at $\pm$ 60° from the minimum principal stress direction. Thus, for the prominent north 60° east direction of the maximum horizontal stress orientation, the conjugates shear zones may have a direction of north 30° east or north 90° east. The north 30°

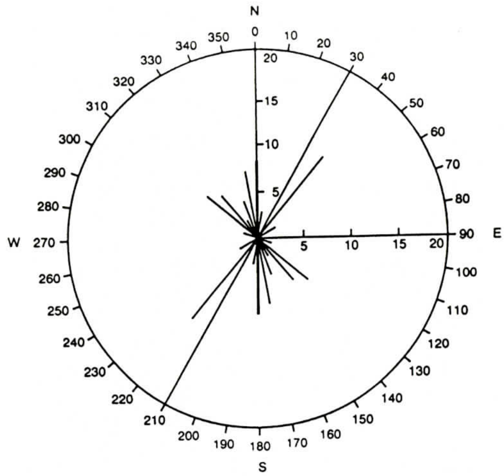


**Figure 14. Frequency distribution of the seismic lineaments given in Figure 13. Note the prominent N30°E orientation of the seismic lineaments. This orientation is in agreement with the direction of the maximum shear stress produced by the stress field assuming a rock frictional angle of 30 degrees.**

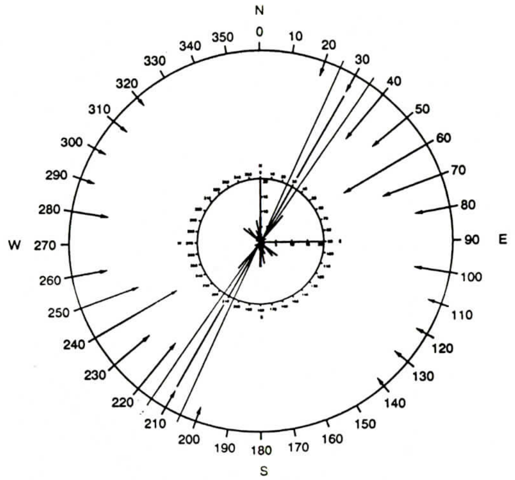


**Figure 15. Azimuthal distribution of the seismic lineaments and stress orientations. The inner circle gives the frequency distribution of the seismic lineaments. The outer circle gives the frequency distribution of the stress orientations. Length of the arrows are proportional to the observed frequencies as in Figure 12. Assuming a frictional angle of 30 degrees, one of the shear plane occurs at N30°E. This is indicated by heavy line on diameter of the larger circle. This direction is in agreement with the observed direction of the maximum number of the seismic lineaments indicated in the smaller circle. The other two thin diagonal lines give the directions of the maximum shear for frictional angle of 20 and 40 degrees.**

east direction is in very good agreement with the observed orientation of the maximum number of seismic lineaments, Figure 15. This agreement was obtained for a frictional angle of 30°, but this angle may vary from 20 to 40°, still disagreement will be only five degrees, this is acceptable considering the large variation in the orientation of the stress axes.

## SUMMARY AND CONCLUSIONS

Seismicity of the southeastern United States is dispersed, and the source mechanisms of the major seismic events are still controversial. In this paper, an attempt is made to correlate seismicity with several possible structures. A majority of the events appear to have a causal relation with the Mesozoic Cenozoic faulting, fold axes, a boundary fault of the Triassic basins, boundaries of the intrusive bodies, post rift dike swarm and magnetic lineaments.Therefore, a preliminary seismotectonics map is proposed. The map shows the probable association of seismic activities with geological structures. Although a majority of the events may be associated with them, still there are several events where no surface structure is mapped or subsurface structure is known. Thus, they cannot be associated with any structure.

An analysis of regional stress field data for the southeastern region indicates a dominant direction of N60°E. This is in good agreement with the directions obtained from recent focal mechanism solutions, (Bollinger and others, 1991; Talwani, 1991; and Madabhushi and Talwani, 1993). The trends of the seismic lineaments vary from N310° to N40° E. The dominant direction is N30° E. This direction is in harmony with the dominant N60° E direction of the regional stresses assuming 30° as rock frictional angle. Based on distribution of the reverse faults and age of their displacements, Prowell (1988) concluded that the regional compression has existed in the crust from early Cretaceous time through the Pleistocene. The origin of this stress distribution may be explained by the ridge-push model or the basal-drag force model for the relative motion of the North American plate with respect to the African plate as proposed by Reding (1984).

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# PALEOPEDOLOGICAL EVIDENCE FOR A EUSTATIC MISSISSIPPIAN-PENNSYLVANIAN (MID-CARBONIFEROUS) UNCONFORMITY IN SOUTHERN WEST VIRGINIA

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## ABSTRACT

**A unit of paleopedogenically altered strata overlies the marine Bramwell Member of the Bluestone Formation (upper Mississippian) in southern West Virginia. This unit, herein named the "Green Valley paleosol complex," comprises the basal portion of the upper member of the Bluestone Formation, and ranges in thickness from 2.1 to 9.3 m. The Green Valley paleosol complex typically lacks internal stratification and exhibits various pedogenic features including root traces, calcareous nodules, pedogenic slickensides, ferruginous mottles and nodules, and distinct horizons. Biostratigraphic constraints indicate an uppermost Mississippian (upper Chesterian) age for the paleosol complex. A pre-Pennsylvanian episode of subaerial erosion apparently truncated the Green Valley paleosol complex at many localities. Contrary to the commonly accepted model of continuous deposition across the Mississippian-Pennsylvanian boundary in southern West Virginia, the Green Valley paleosol complex indicates the existence of a regional unconformity that resulted from prolonged emergence and nondeposition. The Green Valley paleosol complex developed during the global mid-Carboniferous lowstand and probably is related to that event.**

## INTRODUCTION

Although numerous studies have demonstrated that the Mississippian-Pennsylvanian boundary is disconformable throughout most of the Appalachian basin (Dennison, 1983; Englund and Henry, 1984; Rice, 1984; Chesnut, 1988; Ettensohn and Chesnut, 1989; Englund and Thomas, 1990; Beuthin, 1994), the origin and nature of this boundary in the mid-Carboniferous depocenter of southern West Virginia remains controversial. The upper Mississippian-lower Pennsylvanian succession in southern West Virginia is more complete in southern West Virginia than any other area in the central Appalachian basin. Because of this, Englund (1979) and Englund and others (1981, 1983) have inferred depositional continuity across the Mississippian-Pennsylvanian boundary in southern West Virginia. This interpretation is widely accepted and has influenced subsequent studies by Cecil and others (1985), Ettensohn and Chesnut (1989), Englund and Thomas (1990), Chesnut (1992), Henry and Gordon (1992), and Ettensohn (1994). In contrast to the prevailing view, Rice (1985, 1986) and Beuthin (1994) have hypothesized that the Mississippian-Pennsylvanian contact is unconformable even in southern West Virginia.

The nature of the Mississippian-Pennsylvanian boundary in southern West Virginia is crucial to understanding the broader issue of basin-scale controls on mid-Carboniferous sedimentation. A mid-Carboniferous unconformity has long been recognized in many basins that has been attributed to eustatic sea-level change (Sloss, 1963; Vail and others, 1977; Ross and Ross, 1985). Saunders and Ramsbottom (1986) have synthesized biostratigraphic evidence for a global mid-Carboniferous hiatus that comprises the upper Arnsbergian through Alportian stages (uppermost Mississippian) and locally extends into the Kinderscoutian stage (lowermost Pennsylvanian) (Figure 1). This eustatic unconformity includes the widely recognized Mississippian-Pennsylvanian unconformity of the North American craton, and appears on the Exxon sea-level chart as a second-order sequence boundary (Vail and others, 1977). Although eustasy has been recognized as the cause of the unconformity in most basins, many authors have argued that intrabasinal tectonics, rather than eustasy, is the major control on the Mississippian-Pennsylvanian unconformity in the central Appalachian basin (Quinlan and Beaumont, 1984; Tankard, 1986; Ettensohn and Chesnut, 1989; Chesnut, 1990; Willard and Klein, 1990; Ettensohn, 1994). Beuthin (1994), however, has attributed incision of sub-Pennsylvanian paleovalleys in western West Virginia to the mid-Carboniferous eustatic event, and has suggested that an apparent Chokierian-Alportian hiatus in southern West Virginia resulted from the eustatic event.

| SYSTEM (USA) | STAGE (BRITISH ISLES) |
|---|---|
| lower Pennsylvanian (part) | Marsdenian |
| | Kinderscoutian |
| upper Mississippian (part) | Alportian |
| | Chokierian |
| | Arnsbergian |

**Figure 1. Mid-Carboniferous chronostratigraphic scheme. Base of the Pennsylvanian System in southern West Virginia correlates with base of the Kinderscoutian Stage of Europe (Pfefferkorn and Gillespie, 1981, 1982; Englund and others, 1983). Upper Arnsbergian through Alportian strata are absent in mid-Carboniferous sections on several continents (Ross and Ross, 1985; Saunders and Ramsbottom, 1986).**

This report identifies a paleosol complex (the "Green Valley paleosol complex") that occurs near the Mississippian-Pennsylvanian boundary in southern West Virginia. The Green Valley paleosol complex crops out at several localities that are proximal to the West Virginia reference section for the Mississippian-Pennsylvanian boundary proposed by Englund and others

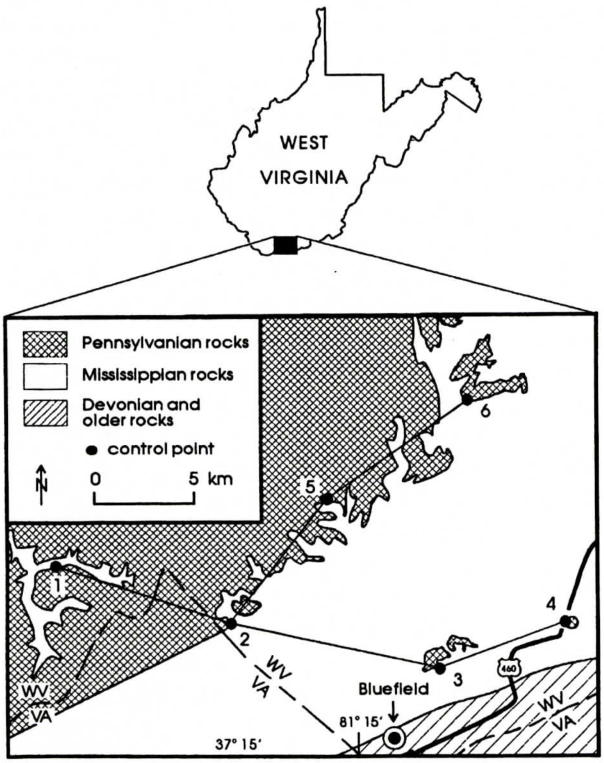


**Figure 2. Location map of sections logged in this study. 1 = Leckie section. 2 = Bluestone section (reference section for the Mississippian-Pennsylvanian boundary in West Virginia). 3 = Route 123 section. 4 = Route 460 section (type section for the Green Valley paleosol complex). 5 = Crystal section. 6 = Lashmeet section. Outcrop pattern based on Englund (1968); Cardwell and others (1968); Trent and Spencer (1990); and personal data.**

(1979) (Figure 2). This paper describes and discusses the Green Valley paleosol complex in the context of the nature and origin of the Mississippian-Pennsylvanian boundary. A preliminary version of this study was presented in Beuthin (1995).

## STRATIGRAPHY AND DEPOSITIONAL MODELS

In southern West Virginia and adjacent parts of Virginia, Englund (1979) has placed the Mississippian-Pennsylvanian boundary within the upper part of the Bluestone Formation (Figure 3). These strata consist primarily of the Mississippian-age Bramwell Member (a medium-dark gray, calcareous, marine fossiliferous siltstone and shale) and the overlying upper member (Pennsylvanian?), which "consists principally of slightly calcareous shale and siltstone that have the typical grayish-red and greenish-gray coloration of the Bluestone Formation (Englund, 1979, p. 72). According to Englund (1979, p. 72) the upper member is Pennsylvanian "because of its intertonguing and lateral grading with the lower sandstone member of the Pocahontas Formation and the presence of *Neuropteris pocahontas*."

At the West Virginia reference section, the systemic boundary is placed at the base of the lower sandstone member of the Pocahontas Formation (Englund and others, 1979) because the upper member of the Bluestone Formation is missing and the sandstone rests directly on the Bramwell Member (Figures 3 and 4). Although the absence of the upper member suggests a hiatus at the systemic boundary, Englund and others (1979, 1981) have interpreted the Bramwell Member and overlying succession of siltstone, shale and sandstone as a prograding delta sequence. According to their depositional model, the Bramwell Member consists of prodeltaic and distal-bar deposits, the lower sandstone member is a delta-front/distributary-channel complex, and the upper member is an interdistributary deposit.

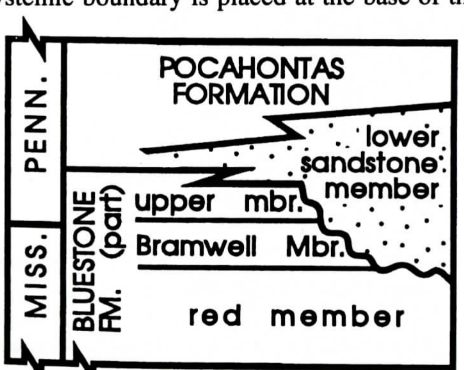


**Figure 3. Lithostratigraphy of the Mississippian-Pennsylvanian boundary, southern West Virginia (after Henry and Gordon, 1992). At the West Virginia reference section, the upper member is absent and the lower sandstone member of the Pocahontas Formation rests directly on the Bramwell Member of the Bluestone Formation.**

About 12 km northeast of the reference section, both the upper member and Bramwell Member are missing, and the lower sandstone member of the Pocahontas Formation rests directly on the red member of the Bluestone Formation (Englund and others, 1981). In accordance with their delta model for late Mississippian-early Pennsylvanian sedimentation, Englund and others (1981, p. 174) have inferred that "scouring in areas near the axes of channels and lobes resulted in the truncation of slightly older sediments. Thus, before the deposition of the lower sandstone member of the Pocahontas, the upper member of the Bluestone, and locally, the Bramwell Member were completely eroded."

Alternatively, Donaldson and Shumaker (1981), Rice (1985), and Rice and Schwietering (1988) have interpreted basal Pennsylvanian strata in southwestern Virginia and southern West Virginia as alluvial plain, rather than delta-distributary deposits. Furthermore, Rice (1985) has postulated that a period of valley incision preceded early Pennsylvanian deposition, and that the Mississippian-Pennsylvanian contact is disconformable.

## THE GREEN VALLEY PALEOSOL COMPLEX AND RELATED STRATA

### General Description and Setting

The Green Valley paleosol complex is a

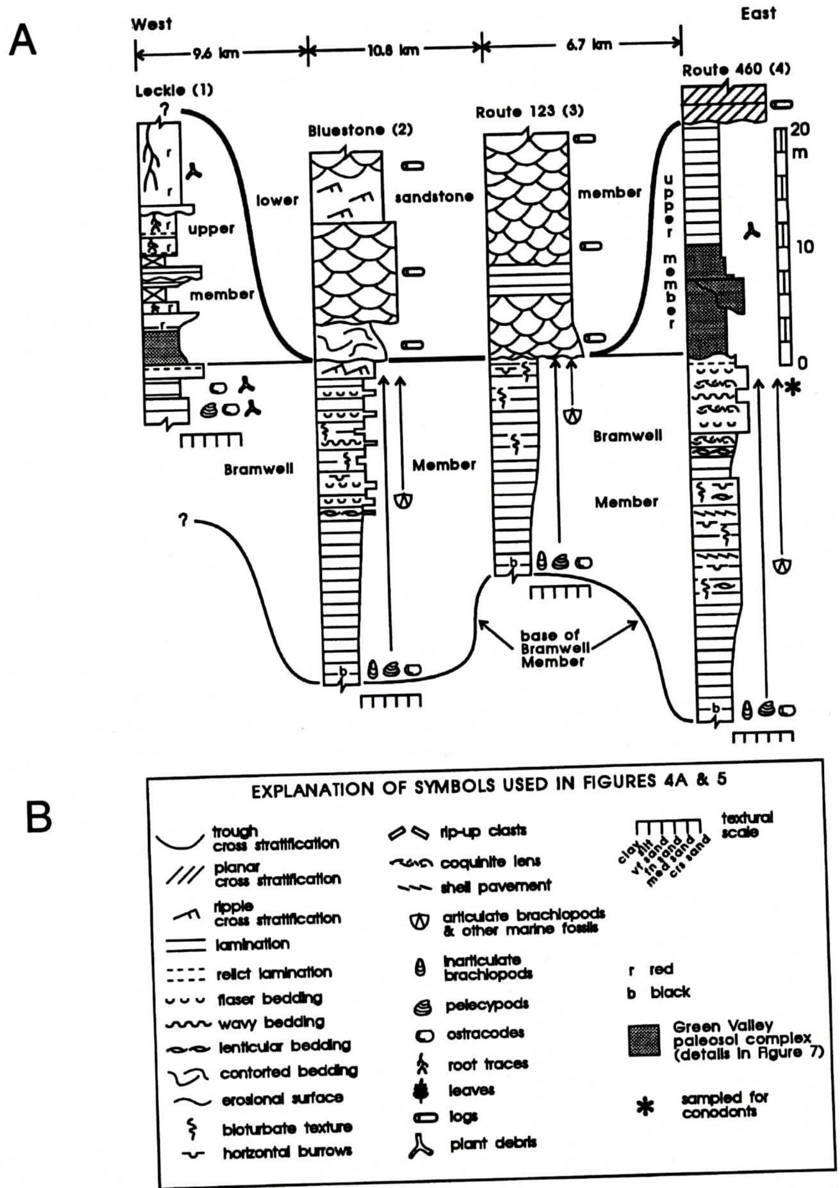


**Figure 4. (A) East-west correlation diagram of mid-Carboniferous successions showing sedimentological and stratigraphic setting of the Green Valley paleosol complex. Datum is Mississippian-Pennsylvanian boundary as defined by Englund (1979). Heavier correlation lines represent Bluestone-Pocahontas contact. Line of section shown on Figure 2. (B) Explanation of symbols used in Figures 4 and 5.**

lithologically distinctive body of rock that directly overlies the Bramwell Member and comprises the basal part of the upper member of the Bluestone Formation (Figures 4 and 5). The paleosol complex is named after the rural community of Green Valley, Mercer County, West Virginia where it is prominently developed and well-exposed (Figure 6). It is referred to as a paleosol complex because at its type section (the Route 460 section) it comprises two superposed paleosol profiles. Red, brown, gray and olive-gray mudstone and claystone are the dominant rock types, but the paleosol complex also includes subordinate siltstone and very-fine-

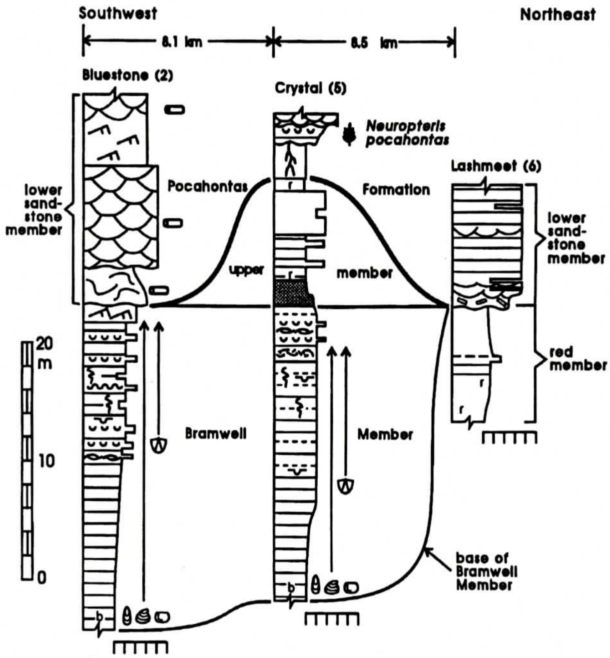


**Figure 5. Northeast-southwest correlation diagram of mid-Carboniferous successions showing sedimentological and stratigraphic setting of the Green Valley paleosol complex. Datum is Mississippian-Pennsylvanian boundary as defined by Englund (1979). Heavier correlation lines represent Bluestone-Pocahontas contact. Line of section shown on Figure 2. See Figure 4 for explanation of symbols.**


**Figure 6. Outcrop of upper part of Bramwell Member (B), upper member of Bluestone Formation (U), and lower part of Pocahontas Formation (P) along U. S. Route 460 in Green Valley, Mercer County, West Virginia. The Green Valley paleosol complex (G) overlies the Bramwell Member along an erosional contact that exhibits about 3 m of relief. Note the massive, internally unstratified character of the Green Valley paleosol complex as compared to the distinctly stratified overlying and underlying rocks. Also, planar cross-bedding is faintly visible in the Pocahontas sandstone at the top of the exposure (dip is to the left, indicating northerly paleocurrents). Contacts are dotted where concealed. Geologist in lower left portion of photo is about 1.8 m tall.**

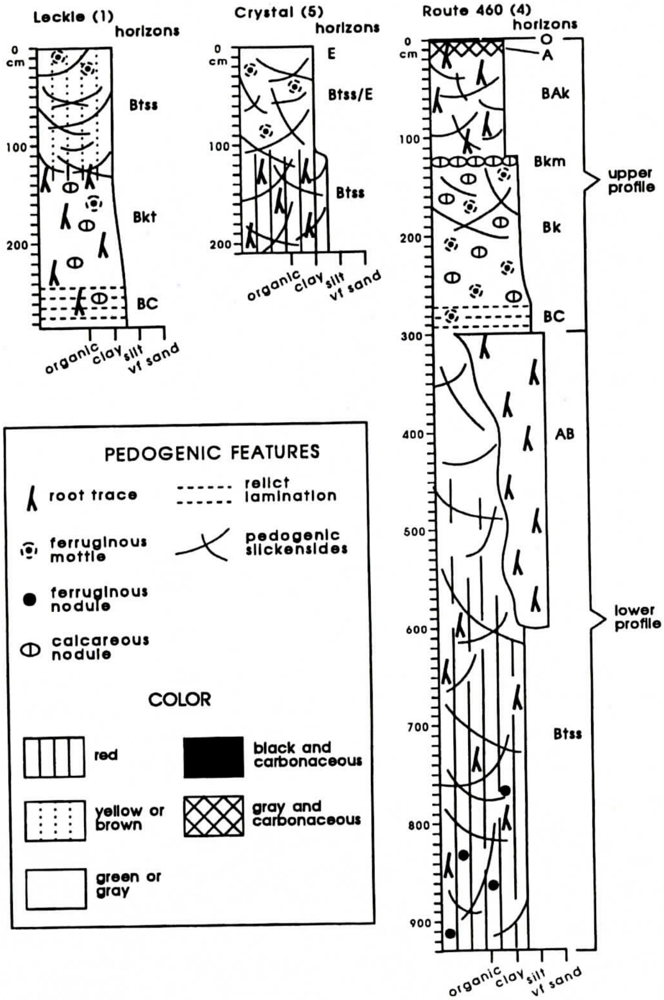


**Figure 7. Graphic logs of the Green Valley paleosol complex. Shorthand for paleosol horizons from Soil Survey Staff (1994). Designation of horizons is based on field criteria.**

grained, argillaceous sandstone. The upper contact is sharp and easily distinguished because the Green Valley paleosol complex generally lacks internal lamination and primary sedimentary structures, whereas overlying strata are distinctly laminated. Thickness of the paleosol complex ranges from 2.1 to 9.3 m.

The Green Valley paleosol complex is known from only three localities in the study area. Despite this, outcrops of the paleosol are situated to the west, north, and east of the reference section for the Mississippian-Pennsylvanian boundary (see Figures 2, 4, and 5). At localities where the Green Valley paleosol is absent, the lower sandstone member directly overlies either the Bramwell Member or the red member of the Bluestone Formation along an erosional contact (Figures 4 and 5).

## Characteristics of Underlying and Overlying Strata

As noted by Englund and others (1981), the Bramwell Member is a coarsening-upward unit that grades upward from laminated clay-shale to flaser- to lenticular-bedded coarse siltstone and very-fine-grained sandstone (Figures 4 and 5). Marine fossils, including articulate brachiopods, bryozoans, pelmatozoans, pelecypods, trilobites, and nautiloids, are abundant in the


**Figure 8. Exposure of part of the BAk horizon in the upper profile of the Green Valley paleosol complex at the Route 460 site. Note calcareous rhizoconcretions in claystone (outlined by dotted lines, and indicated with "R"). Pocket knife is 9 cm long.**

middle part of the Bramwell. Most of these remains are unabraded and are concentrated in shell pavements along bedding planes. The middle part of the Bramwell is also moderately to strongly bioturbated. Marine fossils occur in the upper, coarse-grained part of the Bramwell; however, these fossils are typically abraded and are concentrated in lenses of shell hash. At the Leckie site, in the western part of the study area, the upper part of the Bramwell lacks an open marine fauna. The dominant fossil remains there are plant debris, ostracodes, and scattered pelecypods (Figure 4). Throughout the area, the upper 1 to 2 m of the Bramwell exhibit relict lamination, and appear to be deeply weathered.

The portion of the Bluestone Formation that overlies the Green Valley paleosol complex consists of shale and mudstone with some interbedded, very-fine-grained, argillaceous sandstone (Figures 4 and 5). These strata lack invertebrate fossils, and have yielded only scattered plant debris. At the Leckie site, many of these beds are rooted.

## Paleosol Profiles

In the Route 460 section, the Green Valley paleosol complex consists of two profiles (Figure 7). The 3.0-m-thick upper profile comprises six horizons: O, A, BAk, Bkm, Bk, and BC (in descending order). A 4-cm-thick carbonaceous shale to bony coal at the top of the profile represents the O horizon. The A horizon below this is in a 20-cm-thick, dark-gray claystone containing carbonized root traces and disseminated organic matter. The BAk horizon is 1.0 thick, an is in a medium-dark-gray claystone. The claystone contains pedogenic slickensides, botryoidal calcareous nodules, and calcareous rhizoconcretions cored by carbonized roots (Figure 8). This horizon forms the transition between the A horizon above and the Bkm horizon below. An 8- to 10-cm-thick, medium-dark-gray, argillaceous limestone underlies the BAk horizon. This limestone, which consists of coalesced pedogenic calcareous nodules, forms the Bkm horizon (Figure 9). The Bk horizon is in a 1.4-m-thick mudstone that changes downward from medium-dark-gray to medium-light-gray. Calcareous nodules are common throughout this horizon and pedogenic slickensides occur in the top portion (Figure 8). Root traces are conspicuously absent in the Bk horizon. The basal BC horizon is 0.3 m thick. This horizon is dominantly in a light-olive-gray, clayey siltstone with faint relict bedding. Light-brown to dark-yellowish-orange mottles are scattered throughout this horizon.

The lower profile in the Route 460 section is 6.3 m thick and consists of an AB horizon and a Btss horizon (Figure 7). The AB horizon is most prominently developed in a 0- to 3.0-m-thick, olive-gray, muddy sandstone that contains numerous, downward branching, carbonized root traces, and disseminated organic matter. The sandstone has a convex-downward base and pinches out laterally. Where this sandstone is absent, the AB horizon is unrecognizable on the basis of field criteria. The Btss horizon is in a 3.3- to 6.3-m-thick, clayey mudstone that lacks any trace of relict bedding. Large, wedge-shaped peds bounded by intersecting pedogenic slickensides are distributed throughout the claystone (Figure 10). The upper part of the Btss horizon is olive-gray with light-brown to grayish-red mottles, but the lower part is grayish-red with olive-gray mottles. The red-to-gray color transition is vertically interfingering. Downward branching, drab-haloed root

**Figure 9. Exposure of the Bkm horizon in the upper profile of the Green Valley paleosol complex at the Route 460 site. Pocket knife is 9 cm long.**

traces, some of which are cored by carbonized root hairs, are evident in the red portion of the horizon. Spherical, red, ferruginous nodules ranging from 3 to 8 mm in diameter are also present in the red part of the Btss horizon.

The Green Valley paleosol complex in the Crystal section comprises three horizons: E, Btss/E, and Btss (Figure 7). The E horizon ranges from 0 to 9 cm thick and is developed in a weakly consolidated, lenticular, very-light-gray to dusky-yellow clay. The underlying Btss/E horizon is a 101-cm-thick, combination horizon that consists of two distinct parts. The Btss part is developed in a yellowish-gray to light-olive-gray, slickensided claystone with ferruginous mottles. The E part is represented by discrete lenses of weakly consolidated, very-light-gray clay that are scattered throughout the claystone. The Btss horizon at the base of the profile is 100 cm thick and is developed in a grayish-red mudstone. Slickensides, drab-haloed root traces, and light-olive-brown ferruginous mottles are common throughout the Btss horizon.

The Green Valley paleosol complex in the Leckie section comprises a single profile consisting of three horizons: Btss, Bkt, BC (Figure 7). The Btss horizon at the top of the profile is developed in a 135-cm-thick, dominantly grayish-brown to dusky-brown claystone that becomes grayish-red in the basal 15 cm. Slickensides are prominent throughout the Btss horizon, and drab haloed-root traces are evident in the basal 15 cm of the horizon. The underlying Bkt horizon is 110 cm thick and is devel-

**Figure 10. Exposure of the middle part of the Btss horizon in the lower profile of the Green Valley paleosol complex at the Route 460 site. Note criss-crossing pedogenic slickensides (highlighted with dashed lines) that create large, wedge-shaped peds. Note also the total absence of stratification. Pocket knife is 9 cm long.**

oped in a light-olive-gray mudstone that is clayey in the upper few cm. Calcareous nodules and calcareous rhizoconcretions are abundant throughout the horizon, and grayish-red ferruginous mottles occur in the top 40 cm. In contrast to the overlying Btss horizon, slickensides are not apparent in the Bkt horizon. The 40-cm-thick BC horizon at the base of the profile exhibits the same features as the Bkt horizon, but is distinguished by the presence of faint relict lamination.

## Age of the Paleosol Complex

The Green Valley paleosol complex lacks biostratigraphically useful fossils. However, index fossils contained in underlying and overlying strata set lower and upper age-limits for the paleosol. Conodonts recovered from the Bramwell Member of the Bluestone Formation are reported to belong to the *Adetognathus unicornus* to lower *Rhachistognathus muricatus* zones of Upper Mississippian (upper Chesterian) age (Repetski and Henry, 1983). Samples collected from about 2 m below the top of the Bramwell Member at the Route 460 section contain the conodont *Gnathodus postbileneatus* (Figure 4), which suggests an age no older than uppermost Arnsbergian but no younger than lower Chokierian of the British standard section (see Figure 1) (Robert Stamm, per. comm., 1994). Samples collected from about 11 m above the top of the Green Valley paleosol complex at the Crystal section contain the plant fossil *Neuropteris pocahontas* (Figure 5). *Neuropteris pocahontas* is the main guide fossil for lower Pennsylvanian rocks in southern West Virginia, and its stratigraphic range extends no lower than the base of the Namurian B Series (base of the Kinderscoutian stage) of the British standard section (Pfefferkorn and Gillespie, 1981; Englund and others, 1983).

Available evidence supports an Upper Mississippian (upper Chesterian) age for the Green Valley paleosol. In terms of the British standard section, the Green Valley paleosol complex is post-Arnsbergian, but pre-Kinderscoutian in age.

## DISCUSSION

The Green Valley paleosol complex indicates that soil-forming environments existed in southern West Virginia during the end of the Mississippian Period. Development of the Green Valley complex apparently involved several pedogenic processes. Slickensides suggest alternating expansion and contraction of the soil mass as a result of seasonal wetting and drying, as occurs in modern Vertisols (Wilding and Tessier, 1988). This interpretation is consistent with studies by Cecil and others (1985) and Cecil (1990) that indicate a wet-dry climate during late Mississippian time. Carbonate precipitation was also an important process, resulting in the development of calcareous nodules. The presence of clayey subsurface (Bt) horizons suggests accumulation of clay as a result of illuviation, *in situ* weathering of alumino-silicate minerals, or both. Also, the general absence of stratification and the distinct presence of soil horizons indicate that soil-forming processes thoroughly re-organized the sedimentary parent material. The thick B horizons (>1 m) in the Green Valley paleosol complex indicate a very strong stage of pedogenic development, according to the paleosol development scheme of Retallack (1988). Although the time factor in paleosol development is difficult to quantify, Retallack (1988, p. 16) has indicated that very strongly developed paleosols are "mostly found at major geological unconformities."

Englund (1979) and Englund and others (1981) have inferred that the stratigraphic interval including the Mississippian-Pennsylvanian boundary in southern West Virginia is depositionally continuous. The Green Valley paleosol complex, however, indicates that deposition was interrupted by a late Mississippian episode of emergence. The paleosol complex represents a hiatus that resulted from relatively protracted subaerial exposure and pedogenesis. Although Englund and others (1981) have inferred that the upper member of the Bluestone Formation was deposited in a marshy interdistributary environment, it is unlikely that the Green Valley paleosol complex was developed in such an environment. Coastal marshes are characteristically poorly drained, and coastal-marsh soils typically bear evidence of waterlogging. The Green Valley paleosol complex exhibits deeply penetrating root traces, maturely developed clayey subsurface (Bt) horizons, and calcareous soil nodules. These features are characteristic of well-drained soils, but not of waterlogged soils (Retallack, 1990). Also, Cecil and Dulong (1992, p. 45) have noted that strongly developed paleosols with thick profiles in the red member of the Bluestone Formation "appear to be the result of allocyclic lowering of the water table and pedogenesis on a regional scale." Whereas the thick paleosols in the red member are formed on fluvial deposits (Cecil and Dulong, 1992), the Green Valley paleosol complex is formed on marine sediments of the Bramwell Member. Thus, the Green Valley paleosol complex not only suggests prolonged, regional-scale pedogenesis, but it also suggests a punctuated mid-Carboniferous record controlled by relative sea-level change.

The absence of the Green Valley paleosol complex at many localities may be the result of pre-Pocahontas erosional truncation, rather than facies change. Although Englund (1979) and Englund and others (1981) have inferred an intertonguing relationship between the upper member of the Bluestone Formation and the lower sandstone member of the Pocahontas Formation, evidence for this relationship is lacking. On the other hand, Englund and others (1981) showed that the lower sandstone member locally truncates the upper member and the Bramwell Member; the present study verifies this relationship. Also, at those localities where the lower sandstone member rests directly on the Bramwell Member, the contact is sharp and erosional. These observations suggest that an episode of widespread erosion and local channel incision preceded deposition of the Pocahontas Formation. Rice (1985) has postulated that southern West Virginia and southwestern Virginia underwent a period of fluvial dissection during late Mississippian-early Pennsylvanian time. The erosion surface at the base of the lower sandstone member might be related to this base-level event. Regardless, the Green Valley paleosol complex, and the upper member as a whole, are missing at the reference section for the Mississippian-Pennsylvanian boundary and nearby localities because they were truncated during an episode of pre-Pocahontas subaerial erosion.

Hypothetically, the Green Valley paleosol complex could record a period of tectonic uplift in southern West Virginia, but this seems unlikely. According to several studies, southern West Virginia was a late Mississippian-early Pennsylvanian depocenter where subsidence was relatively continuous because of thrust-loading on the eastern orogenic margin of the central Appalachian basin (Quinlan and Beaumont, 1984; Tankard, 1986; Ettensohn and Chesnut, 1989; Ettensohn, 1994). Furthermore, these studies have treated the Mississippian-Pennsylvanian unconformity on the western flank of the basin as a product of early Pennsylvanian flexural uplift that occurred after deposition of the Pocahontas Formation. The late Mississippian Green Valley paleosol complex clearly originated prior to the early Pennsylvanian uplift that produced the Mississippian-Pennsylvanian disconformity on the basin flank. Apparently, the Green Valley paleosol complex represents a hiatal surface that is truncated toward the basin margin by the tectonically-controlled early Pennsylvanian surface.

Although the general episode of emergence recorded by the Green Valley paleosol complex does not appear to be a tectonically controlled, it probably is eustatically controlled. The interregional hiatus that records the mid-Carboniferous eustatic event includes the upper Arnsbergian, the Chokierian, and the Alportian Stages, and spans a time interval of as much as 4.5 Ma (Saunders and Ramsbottom, 1986). The general post-Bramwell regression recorded by the Green Valley paleosol complex therefore corresponds to the mid-Carboniferous global lowstand. The presence of two profiles in the Green Valley paleosol complex at the Route 460 section indicates that deposition occurred at least locally during the mid-Carboniferous eustatic event. This perhaps resulted from variations in the rate of subsidence in the southeastern part of the study area. If, for a period of time, the rate of subsidence exceeded the rate of eustatic fall, then relative sea-level rise would have occurred. This could have resulted in submergence, deposition, and ultimately burial of the older profile at the Route 460 section. Later, when then rate of subsidence slowed, relative sea level again dropped. This, in turn, resulted in renewed emergence and pedogenesis at the Route 460 site. Although a detailed account remains conjectural, there appears to be a general cause-and-effect relationship between the mid-Carboniferous eustatic event and the formation of the Green Valley paleosol complex. Ongoing subsidence in southern West Virginia probably mitigated the effect of the eustatic drop on fluvial base level; thus, the eustatic event resulted in prolonged exposure and pedogenesis, rather than deep incision and erosional beveling.

It is also possible, but less certain, that the erosional contact between the Bluestone and Pocahontas formations is a product of the mid-Carboniferous eustatic event. Mid-Carboniferous plant megafossils from southern West Virginia record a significant floral break at the Mississippian-Pennsylvanian boundary (Pfefferkorn and Gillespie, 1981, 1982) that seemingly contradicts reports of depositional continuity (Englund, 1979; Englund and others, 1981, 1983). Furthermore, on the basis of this floral break, Pfefferkorn and Gillespie (1981, 1982) correlated the Mississippian-Pennsylvanian boundary with the Namurian A-B boundary of Europe (Alportian-Kinderscoutian boundary as shown in Figure 1 of the present study). Existing biostratigraphic control suggests that pre-Pocahontas erosion occurred no later than Kinderscoutian time. In several basins, the mid-Carboniferous eustatic event lasted into Kinderscoutian (very early Pennsylvanian) time (Saunders and Ramsbottom, 1986). Therefore, the general timing of the pre-Pocahontas erosion appears to correspond to the later part of the mid-Carboniferous eustatic event. As with the Green Valley paleosol complex, the episode of pre-Pocahontas erosion perhaps resulted from a high rate of eustatic fall relative to the rate of subsidence in southern West Virginia.

## CONCLUSIONS

Although the Mississippian-Pennsylvanian boundary succession in southern West Virginia is more complete than in any other area in the central Appalachian basin, it seems unlikely that continuous deposition occurred during latest Mississippian-earliest Pennsylvanian time. The Green Valley paleosol complex contains thick paleosol profiles at the top of the Mississippian section, which argue for prolonged periods of exposure, and an interruption in the previously interpreted facies succession. In addition, basal Pennsylvanian strata above the paleosol profiles are sharp-based with no evidence of lateral interfingering. Locally, these sandstones truncate the Green Valley paleosol complex, resulting in a stratigraphic section that may appear to be depositionally continuous, but is actually disconformable. The lack of sedimentological continuity between the Bluestone and Pocahontas Formations indicates the need for additional study of the sedimentary dynamics and paleogeography of southern West Virginia, particularly in the context of allocyclic controls. Lithostratigraphic, biostratigraphic and sedimentological evidence from the Green Valley paleosol complex and associated strata suggests that the eustatic mid-Carboniferous unconformity, which is so prominent in other basins, also extends into the central Appalachian basin in the area of the West Virginia reference section for the Mississippian-Pennsylvanian boundary.

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# BRAID-DELTA FACIES INTERPRETED FROM CORES, GRANNY CREEK OIL FIELD OF WEST VIRGINIA

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## ABSTRACT

**Based on the analysis of 12 full-bore cores, the Big Injun sandstone (Mississippian Price Formation) in and around Granny Creek oil field is divided into two major facies of a braid delta. We interpret the lower, fine-grained litharenite and sublitharenite to be of a distributary-mouth bar. Subfacies include (1) the mouth-bar front: very fine sandstone, shaly and organic rich, with horizontal and inclined parallel laminae; (2) the bar crest: fine sandstone, texturally mature, slightly pebbly, with small-scale trough cross-laminae; and (3) the back of the mouth bar: an interbedding of fine sandstone with parallel laminae and pebbly coarse sandstone with channel-fill cross-laminae. The bar crest is the best reservoir and the primary pay zone. Porosity ranges from 10 to 23%, and permeability, up to 24 md. Primary pores are relatively small, typically less than 0.100 mm, but well connected and not greatly reduced by cement.**

**The upper Big Injun, consisting of very coarse quartzarenite and sublitharenite, is fluvial in origin. Subfacies include (4) channel lag: conglomerate with a scoured base and large-scale trough cross-laminae; and (5) transverse bar: medium to coarse sandstone with large-scale planar cross-laminae and fining-upward sequence. Calcite and quartz cement significantly reduce porosity in the fluvial section although permeability may be high due to the rocks' large pore throats.**

## INTRODUCTION

Braid deltas comprise coarse-grained channel deposits of braided fluvial distributaries. Clast-supported coarse sandstone and pebbly conglomerate with abundant cross-bedding characterize such deltas (McPherson and others, 1987). These rocks can have excellent reservoir quality (Ethridge and Wescott, 1984; Hamlin and others, 1996) although they have often been misidentified in the geological record. The matrix-free, moderately to well sorted sandstone and conglomerate possess high initial porosities and permeabilities. Stratigraphic traps develop because of the considerable lithologic variability, and reservoirs exhibit significant lateral continuity (McPherson and others, 1987).

The Granny Creek field of Clay and Roane Counties, West Virginia (Figure 1), produces oil from the Mississippian Big Injun sandstone, a stratigraphic unit that displays many of the features common to braid deltas. The Big Injun is an informal drillers' term for a sandstone near the top of the Price Formation. Discovered in 1924, the Granny Creek field contains approximately 700 wells. Initial well production was as great as 70 barrels of oil per day, with an average of 25, but today most are stripper wells producing less than 10 BOPD.

Estimates of the total volume of oil originally in place vary from 17 to 68 million barrels, and perhaps 9.8 million were produced by primary methods (Hohn and others, 1993b). An additional 2.1 million barrels have been produced from the northern half of the field since a waterflooding program began in the 1970s. Large volumes of oil remain in place, yet production

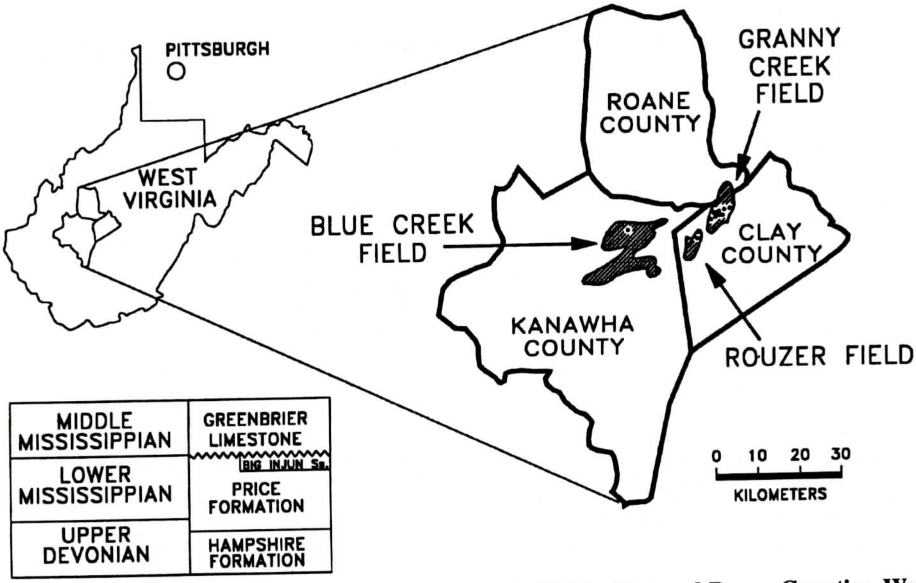


**Figure 1. Map showing the location of Granny Creek oil field in Clay and Roane Counties, West Virginia, nearby Rouzer oil field in Clay County, and Blue Creek oil field in neighboring Kanawha County. Ten of the 12 Big Injun cores (dots) used for this study came from Granny Creek. Inset at the lower left shows the generalized stratigraphy of interest.**

is declining. Full development of the field, especially through methods of secondary and tertiary recovery, is hampered by a poor understanding of the reservoir's stratigraphic framework.

The purpose of this paper is to describe, characterize, and interpret depositional facies of the Big Injun sandstone in the Granny Creek and nearby Blue Creek and Rouzer oil fields. The reservoir's heterogeneity can thus be better understood. Twelve full-bore cores (8 cm in diameter) were analyzed for this study: ten from wells in the Granny Creek field of Clay County (Figure 2), one from a well in the Rouzer field of Clay County, 6 km to the south, and one from a well in the Blue Creek field of Kanawha County, 20 km to the southwest. Wireline logs, whole-core porosity and permeability data, and representative thin sections were also analyzed for this study.

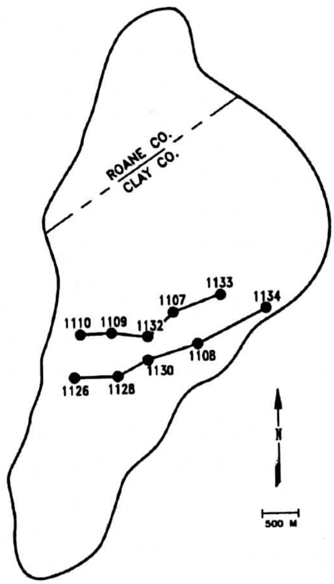


**Figure 2. Distribution of 10 full-bore cores described and interpreted from Granny Creek field. The cross sections of Figures 10 and 13 are also indicated.**

## GEOLOGICAL SETTING

The uppermost Devonian-Lower Mississippian Price Formation reflects the fourth and last tectonic phase of the Acadian orogeny (Ettensohn, 1985). Sediments accumulated as a large deltaic complex (Figure 3), a prograding sandy coastline fed from the east by a few major rivers (Matchen and Kammer, 1994). Following a sea-level highstand in earliest Mississippian time, marginal-marine facies prograded basinward as

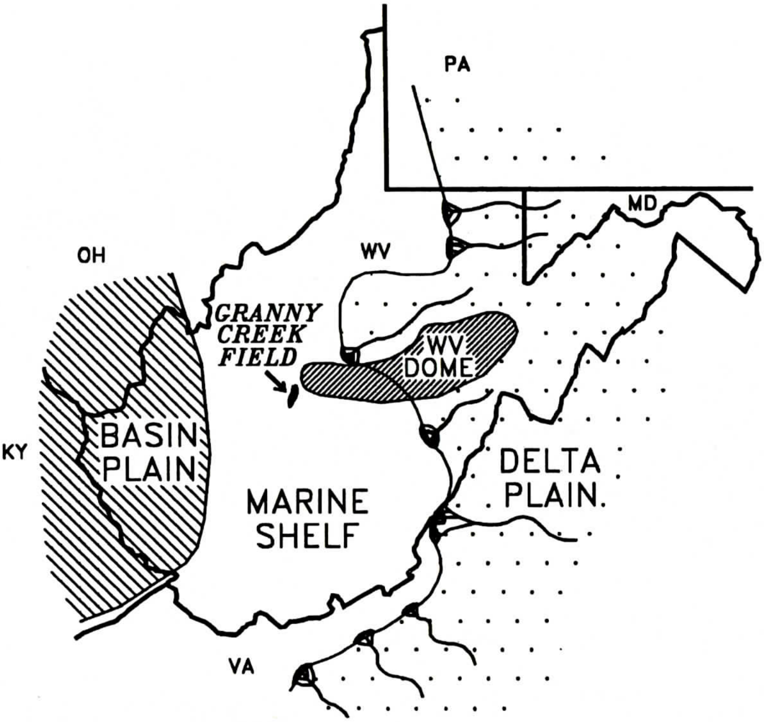


**Figure 3. Paleogeographic reconstruction of the central Appalachian basin during sedimentation of the upper Price Formation (from Bjerstedt and Kammer, 1988). The coastline prograded basinward as two delta systems: a northern river-dominated delta onto a shallow shelf and a southern wave-dominated delta onto a deep shelf. Situated between the two was the West Virginia dome. Granny Creek field lies immediately west of this dome.**

two delta systems (Bjerstedt and Kammer, 1988).

In southern West Virginia and Virginia, the upper section of the Price Formation records the filling of a deep basin by a wave-dominated delta: a coarsening-upward sequence of basinal shale, prodeltaic submarine fan, shoreface sandstone, channel-fill sandstone, and delta-plain coal. By contrast, the upper Price section in northern West Virginia and Maryland contains the fining-upward sequence of a shallow, river-dominated delta: fluvial conglomerate, sandstones of channel fill and crevasse splay, and coastal-plain red mudstone.

Located between the two delta systems was the West Virginia dome, a topographically positive area. The dome was exposed in Early Mississippian time, and erosion locally removed or thinned Price sediments (Donaldson and Shumaker, 1981; Yeilding and Dennison, 1986). Approximately 45 m of the Price Formation were eroded at the unconformity along its crest (Bjerstedt and Kammer, 1988). The source area for coarse-grained siliciclastics encountered in the Granny Creek oil field, situated immediately west of the dome (Figure 3), is thought to be the dome itself. We suggest that very coarse sediment in the middle Price Formation, such as pebbly sandstone and conglomerate of the Rockwell Member, was eroded from the dome's crest and reworked into younger Big Injun facies just to the west and down the regional paleoslope.

Deposition took place on a small braid delta that prograded across the shallow-marine shelf (Figure 4). At Granny Creek the Big Injun consists of two major facies, and each of these is divided into several subfacies. As will be discussed below, we interpret the lower fine-grained sandstone facies to be of a distributary-

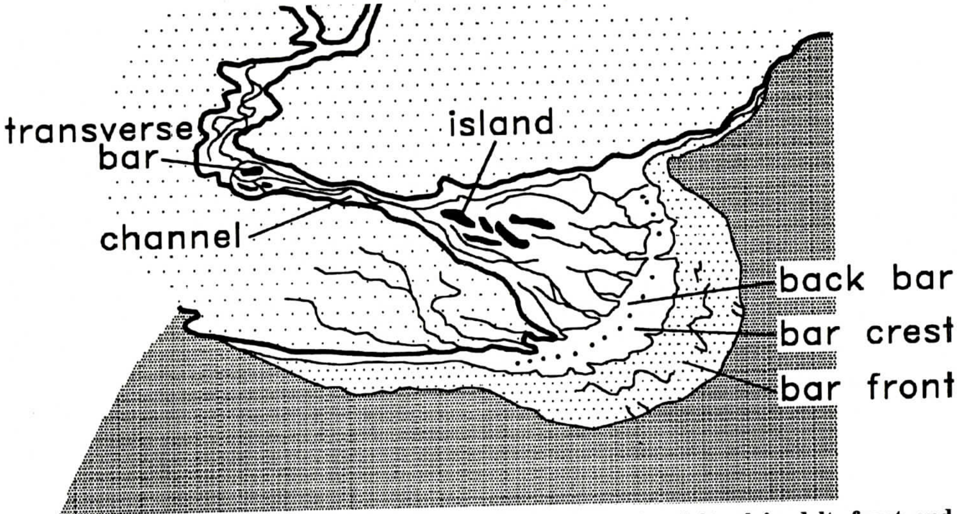


**Figure 4. Sketch of the Price braid delta at Granny Creek, illustrating delta plain, delta front, and the several subfacies.**

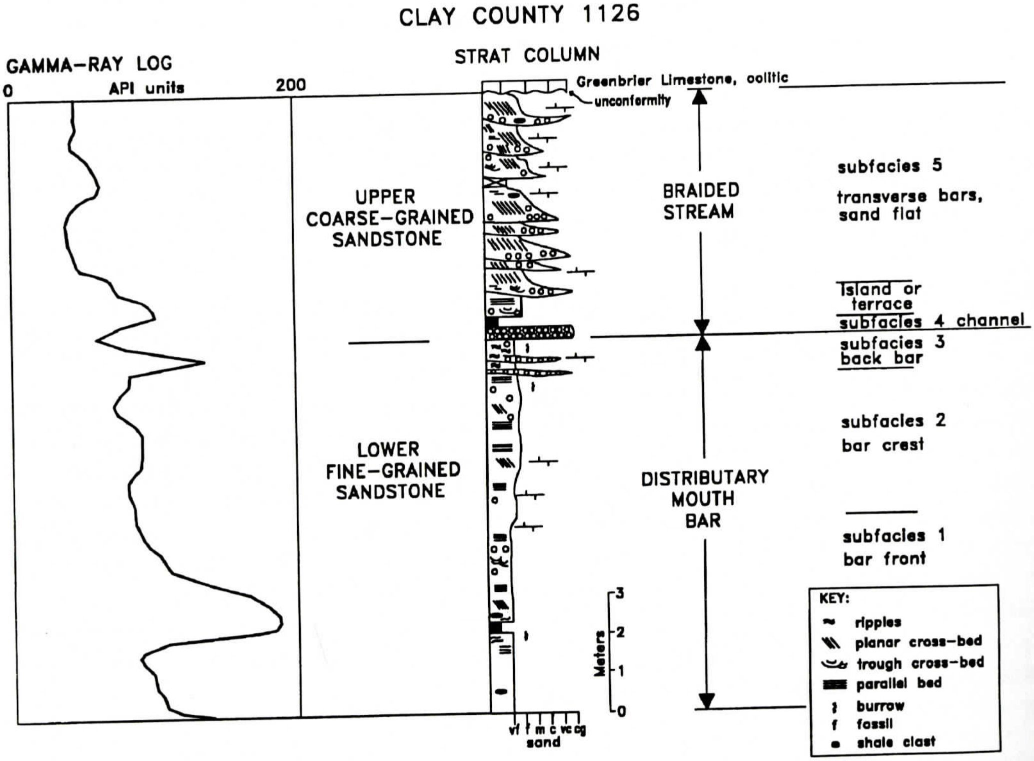


**Figure 5. Stratigraphic section of the Big Injun sandstone in the Clay 1126 core, depicting grain size, important sedimentary structures and other petrographic characteristics, nature of the gamma-ray log, and facies interpretations.**

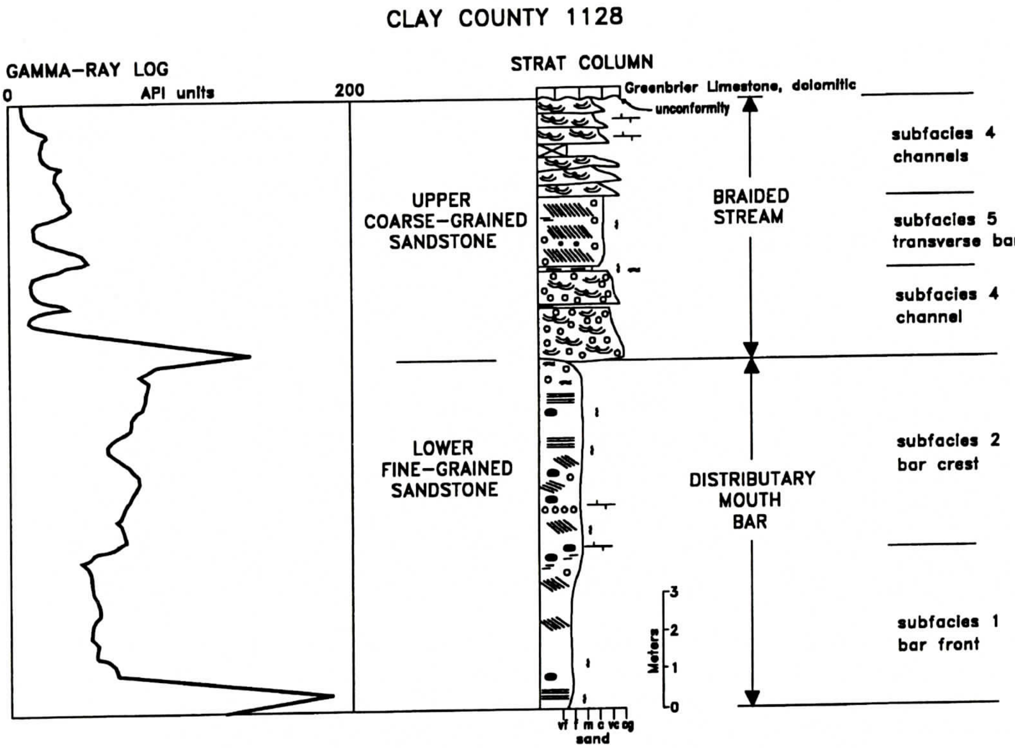


**Figure 6. Stratigraphic section of the Big Injun sandstone in the Clay 1128 core, depicting grain size, important sedimentary structures and other petrographic characteristics, nature of the gamma-ray log, and facies interpretations. Same symbols as used in Figure 5.**

mouth-bar environment on the delta front. The upper coarse-grained sandstone facies is of a fluvial setting on the delta plain. The same facies have been identified at Rouzer oil field.

## LOWER, FINE-GRAINED SANDSTONE

The lower unit of the Big Injun member has been correlated and mapped across the Granny Creek field based on its gamma-ray and bulk-density signatures. It reaches a maximum thickness of 18 m in the southwestern sector but more typically ranges between 5 and 12 m (Hohn and others, 1993b). Within this lower unit Zou (1993) identified three laterally accreting tongues, labelled C1, C2, and C3 and separated from one another by shale. Mapping of the central C2 tongue (the only one completely within the field) shows it to be approximately 9 m in maximum thickness, 2 km wide, and with a north-south trend (Zou, 1993). Presumably the others are of similar dimension and orientation.

All ten cores of the present study penetrated the C2 tongue, providing a clear picture of both vertical and lateral facies changes. In addition, two cores penetrated the western flank of the C1 tongue, and seven cores the eastern half of the C3 tongue. Detailed sedimentologic analysis of these cores has distinguished three subfacies within the several tongues — identified simply as subfacies 1, 2, and 3 (Figures 5 and 6). The subfacies aggraded vertically through time, with 1 at the base and 3 at the top, but they also accreted laterally as the sandstone tongues built both upward and seaward.

### Subfacies 1

The lowest subfacies consists of very fine sandstone with abundant organic matter, mica, and pyrite and with siderite and calcite cements

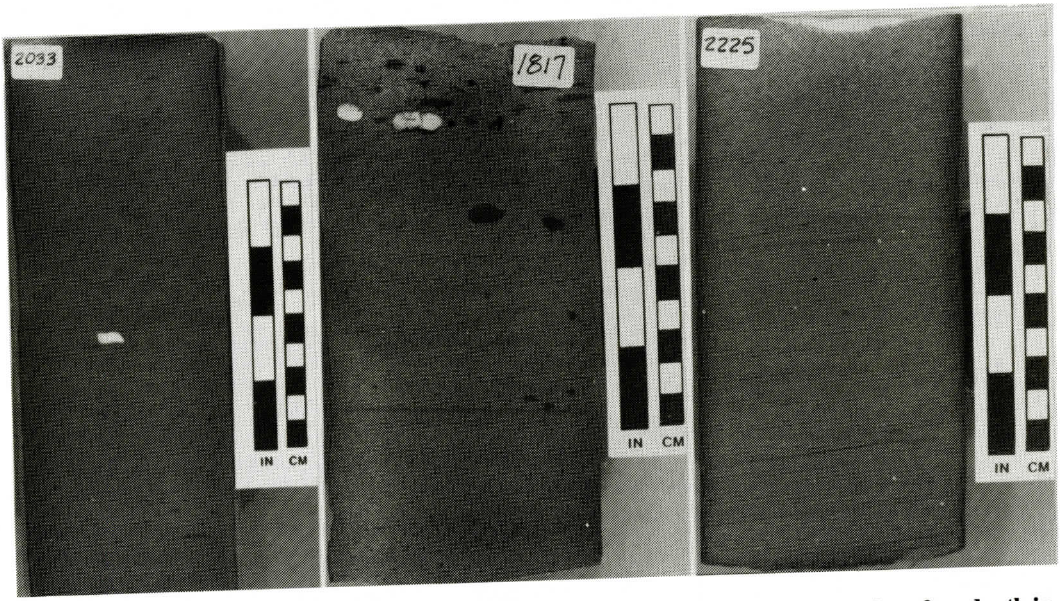


Figure 7. Core photographs of the bar-front subfacies. Number on the cores is the subsurface depth in feet. *2033.* Vaguely laminated to structureless, very fine sandstone with a single floating quartz pebble (10 mm). Clay 1128 core. *1817.* Very fine sandstone with two quartz pebbles and a number of black-shale clasts. Clay 1126 core. *2225.* Sandstone showing inclined parallel laminae at bottom and small-scale cross-laminae in middle. Clay 1134 core.

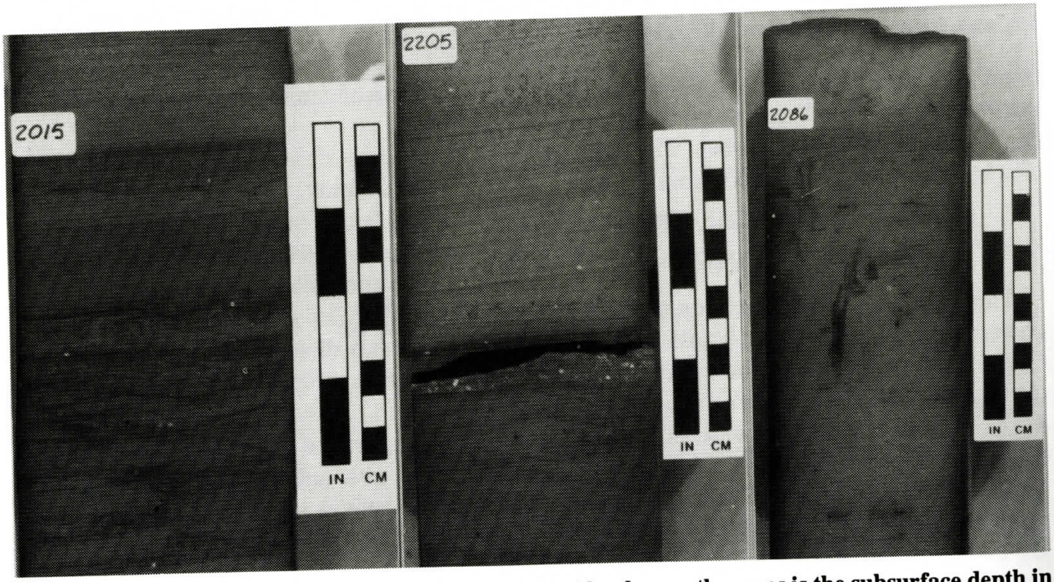


Figure 8. Core photographs of the bar-crest subfacies. Number on the cores is the subsurface depth in feet. *2015.* Upper half of this core shows parallel laminae and one shallow channel-fill structure; lower half shows small-scale trough cross-laminae (ripple bedding) disturbed by a few burrows. Clay 1128 core. *2205.* Very fine sandstone with a thin pebbly layer (in middle at break). Structures include low-angle planar cross-laminae above and small-scale trough cross-laminae (ripple bedding) below. Clay 1134 core. *2086.* Very fine sandstone with vague laminae and several large, vertical and horizontal burrows (with black organic linings). Clay 1108 core.

and diagenetic nodules. Very coarse sand and pebbles of quartz (maximum length of 26 mm, but most less than 5 mm) float in the very fine sandstone or rarely constitute single-grain laminae (Figure 7-2033). The major sedimentary structures include horizontal, parallel lamination and low-angle inclined lamination (Figure 7-1817), but where there is no vertical change in grain size or mineralogy, the rock appears structureless. Small-scale planar and trough cross-laminae are present though uncommon (Figure 7-2225), and a few beds have been burrowed by small infaunal organisms. Toward the bottom of the section are partings and laminae of shale plus eroded clasts of shale, siltstone, and shaly sandstone (up to 4 cm long).

## Subfacies 2

The middle subfacies consists of fine to very fine sandstone. Quartz pebbles and coarse sand are more common in these rocks, floating in the sandstone and occasionally forming layers that range in thickness from 1 mm (single grain) to 3 cm (very thin bed). These layers usually exhibit a scour base. Shale partings and rip-up clasts are rare, and interstitial clay is completely lacking. Texturally the sandstones are mature: clay matrix has been removed, sorting is good, but the grains remain angular. Horizontal, parallel lamination and low-angle inclined lamination are present (Figure 8-2015), as in the underlying subfacies, and concentrations of organic matter, mica, or clay minerals generally mark the bedding planes. Small-scale trough and planar cross-laminae, however, are more common (Figure 8-2205). Not infrequently vertical burrows interrupted these bedding structures (Figure 8-2086). A few pyritized thin-shelled brachiopods are also present.

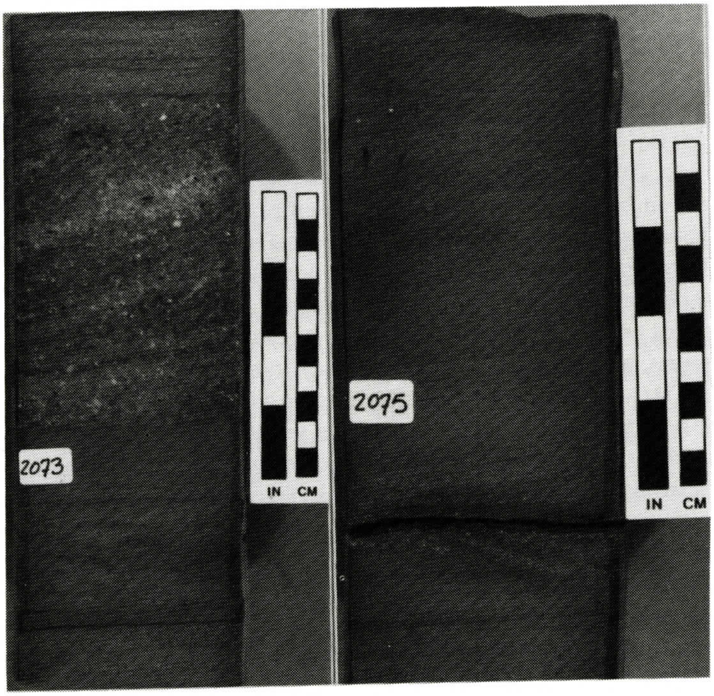


**Figure 9. Core photographs of the back-bar subfacies. Number on the cores is the subsurface depth in feet. *2073.* Fine sandstone with parallel bedding, dark organic-rich laminae, and small-scale cross-laminae (near base, disturbed by escape burrows) alternates with very coarse sandstone with large-scale trough cross-laminae. Clay 1108 core. *2075.* Fine sandstone with low-angle planar cross-laminae alternates with medium to coarse sandstone displaying channel-fill cross-laminae (2074.7 and 2075.1). Clay 1108 core.**

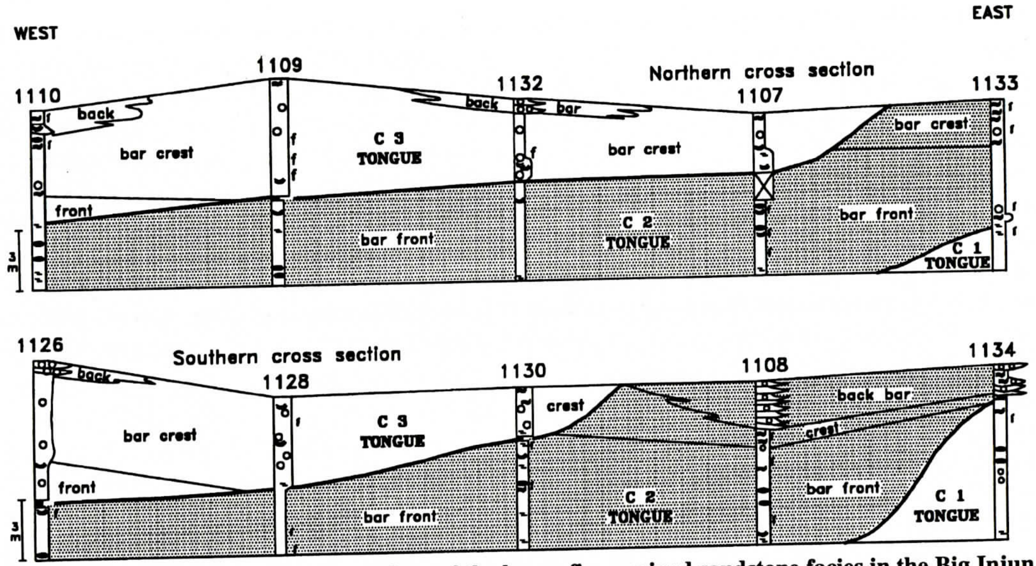


**Figure 10. Two west-to-east cross sections of the lower, fine-grained sandstone facies in the Big Injun member. Only the easternmost two wells penetrated the C1 tongue. The overlying C2 tongue (shaded), which prograded across Granny Creek field, was encountered in all 10 cores. The youngest C3 tongue is identified in the western wells. Shale breaks separate the three sand bodies, and each consists of bar-front, bar-crest, and back-bar subfacies. Lines of section are shown on Figure 2.**

## Subfacies 3

The Big Injun member contains a third subfacies in five cores (#1108, #1110, #1126, #1132, and #1134). This uppermost subfacies, never exceeding 2 m in thickness, consists of two interbedded lithologies (Figure 9-2073). (1) The first is a pebbly, very fine to fine sandstone with parallel lamination and small-scale cross-lamination. Partings of organic-rich shale are common throughout, and burrows (including horizontal *Planolites*) are present locally. (2) The second lithology is a coarse sandstone to conglomerate in which mean grain size ranges up to 10 mm. These coarse-grained rocks occur in discrete beds (as many as 13 in well #1108) from a single grain in thickness up to 12 cm (thin bed). They have a sharp, erosive base and channel-fill cross-lamination (Figure 9-2075). Most beds display a fining-upward sequence: one 11-cm bed, for example, fines upward from a basal very coarse sandstone, to parallel-laminated fine sandstone, to trough cross-laminated fine sandstone with drapes of green micaceous shale.

## Interpretations

Sandstones of the lower Big Injun unit are fine-grained and texturally mature, and grain size coarsens upward from very fine to fine sand (Figures 5 and 6). Bedding structures display a vertical change upward in the cores: horizontal and inclined parallel lamination in the lower section, small-scale trough cross-lamination above, and channel-fill cross-lamination at the very top. Though not common, marine trace fossils and brachiopods have been preserved in the sandstones. Based on this assemblage of sedimentary textures, structures, and fossils (compare with the models of Coleman and Prior, 1982; Elliott, 1986) plus facies associations and the regional setting, the lower fine-grained sandstone is interpreted as a distributary-mouth bar. Supporting evidence from the wireline-log character (Figures 5 and 6; funnel motif of Selley, 1985; Coleman and Prior, 1982) and sand-body geometry (that of a delta with mixed river-wave interaction; Elliott, 1986) are consistent with this interpretation (see Zou, 1993). Although not cored, the underlying 7 to 9 m contain interbedded shale, siltstone, and very fine sandstone, thought to be the offshore prodeltaic mud.

Subfacies 1 is interpreted as a bar-front sandstone, deposited at or above wave base on the seaward-sloping surface of the mouth bar. Parallel lamination was produced when sand settled rapidly from the suspended load at a velocity below the threshold of ripple formation, and shaly sandstone at the bottom of the section marks a downward transition into the prodeltaic facies. Subfacies 2 is interpreted as a bar-crest sandstone, deposited at or near sea level on top of the mouth bar. Trough and planar cross-laminae formed when wave ripples migrated across the surface, and deep-burrowing suspension feeders *Skolithus* and *Cylindrichus* inhabited these mobile sands (Bjerstedt, 1987). Subfacies 3 represents sand and gravel deposits of the back bar, transitional in nature between the crest of the distributary-mouth bar and its feeder stream channel. Waves and currents produced the ripple cross-laminae, whereas interbeds of coarse sandstone and conglomerate represent fluvial sediments transported to the mouth-bar during times of flood discharge.

Architectural makeup of the mouth-bar tongues can be analyzed in detail along the east-west section lines of cores (Figure 10). The C2 sandstone tongue, for example, reaches a maximum thickness of 8.2 m and width of 2.6 km. Initial deposition of the bar-crest and back-bar sandstones occurred just seaward of the underlying tongue's central axis. These upper two C2 subfacies then prograded seaward 1.5 km to a position just west of well #1108, where together they constitute one third of the bar's total thickness. The bar-front subfacies, however, extends 1.6 km farther to the west. The shale break which separates the C2 and C3 tongues exhibits a mean landward rise of 0.38 percent, indicating the depositional slope angle of the C2 bar front (assuming that the base of the Big Injun member was horizontal).

## UPPER, COARSE-GRAINED SANDSTONE

At Granny Creek the upper unit of the Big Injun member has a maximum thickness of 9 m and thins progressively to the northeast (Hohn and others, 1993b). In fact, this unit is absent in the northeast sector of the field. Thinning is due to erosion associated with a regional post-Price unconformity. In the ten cores from Granny Creek, the upper coarse-grained unit thins to the east from 6.4 to 2.4 m. Detailed sedimentologic analysis of these cores has distinguished two subfacies — 4 and 5 (Figures 5 and 6). Subfacies 4 occurs at the base of the unit, but the two then alternate through the overlying stratigraphic section.

### Subfacies 4

This rock consists of poorly sorted, coarse to very coarse sandstone, often pebbly, and locally conglomerate (Figure 11-1795). Well rounded quartz grains range up to 14 mm, and shale clasts, up to 17 mm. Although grain size varies, often dramatically, from lamina to lamina, many beds display a fining-upward sequence, passing vertically to medium sandstone and to fine-grained, silty sandstone. Many beds, too, are capped by a thin, shaly and ripple-laminated sandstone; tiny burrows occur in this green very-fine sandstone. To a lesser degree, illitic and micaceous partings drape bedding planes throughout the subfacies. The matrix between coarse grains consists of fine sand.

Bedsets, which are marked by a sharp erosional contact (Figure 11-1897), range up to 60 cm in thickness but average near 15 cm. Large-scale, trough and planar cross-beds constitute the major sedimentary structures (Figure 11-2069), but the pebbly sandstone frequently appears structureless. In addition, the sandstone may contain small scours or rare slump structure. Elongate grains are oriented down the inclined cross-bedding though they are not imbricate.

### Subfacies 5

The last subfacies contains medium-to coarse-grained sandstone in beds 0.2 to 1.7 m thick. The sandstone base is generally sharp and locally erosive (Figure 12-1787). Grain size may coarsen upward from medium to coarse

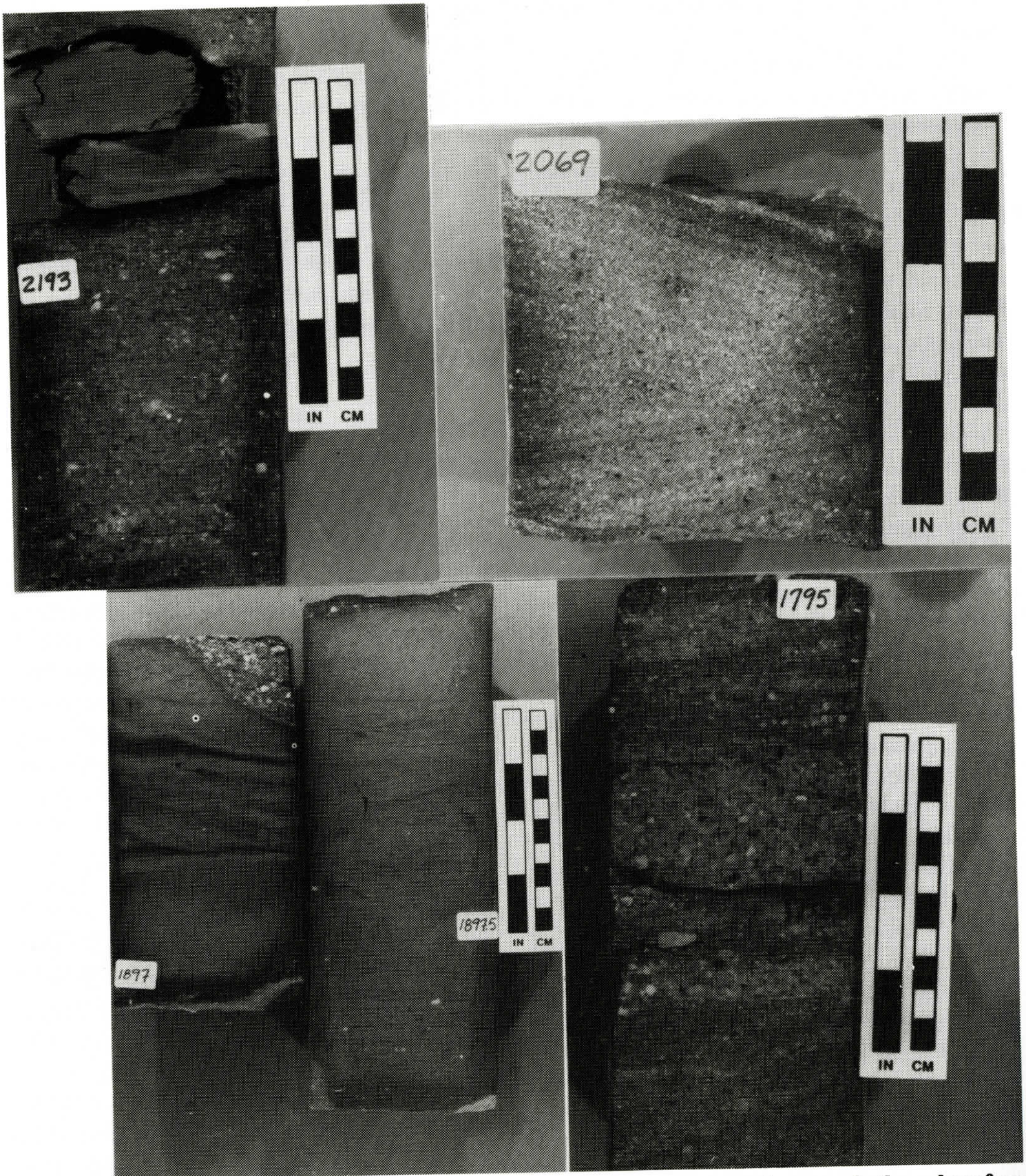


**Figure 11. Core photographs of the channel-floor subfacies. Number on the cores is the subsurface depth in feet. *2193.* Pebbly coarse sandstone with vague cross-laminae is interbedded with green shale (highly fractured). Clay 1134 core. *2069.* Coarse sandstone with large-scale planar cross-bedding. Clay 1108 core. *1897.* Erosive base of pebbly very coarse sandstone has scoured into cross-laminated, burrowed fine sandstone of the underlying bar-crest subfacies. Clay 1109 core. *1795.* Channel-lag conglomerate, having well rounded clasts up to 13 mm, is interbedded with cross-bedded coarse sandstone. Clay 1126 core.**

sand, or more frequently may fine upward from very coarse/coarse to medium/fine/very fine sand (Figure 12-2002). Rare pebbles are up to 10 mm in diameter. Sorting is a function of grain size: coarser sandstone contains admixed pebbles and is poorly sorted, but finer sandstone exhibits good sorting.

The predominant sedimentary structure

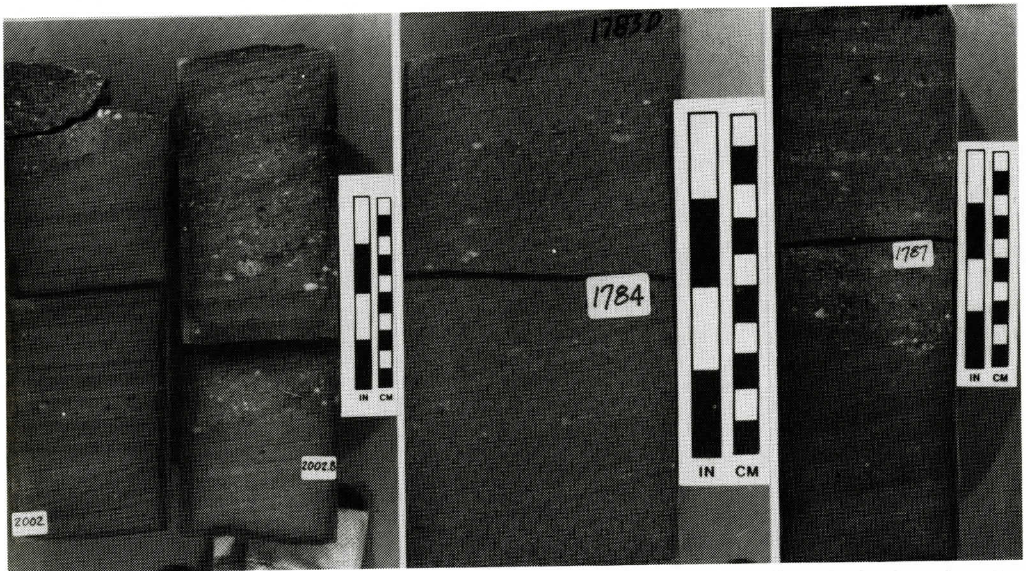


**Figure 12. Core photographs of the transverse-bar subfacies. Number on the cores is the subsurface depth in feet. *2002.* A number of stacked, fining-upward sandstone sequences (one at upper left and two to the right, above and below the break). In each, a sharp basal contact is overlain by pebbly very coarse sandstone with high-angle cross-bedding. This grades upward to medium sandstone with low-angle cross-laminae and/or parallel laminae highlighted by green-shale partings. Clay 1128 core. *1784.* Fine sandstone with floating pebbles up to 7 mm; grain size varies from one lamina to another. Unit displays large-scale planar cross-laminae (14°). Clay 1126 core. *1787.* Pebbly very coarse sandstone has scoured into a fine sandstone with planar cross-laminae. The relief is 2 cm. This unit, one of several cycles, grades upward to an alternating, medium/very coarse sandstone. Clay 1126 core.**

(usually highlighted by partings of organic matter, clay, or mica) is large-scale planar cross-lamination, with beds inclined between 4° and 22° (Figure 12-1784). Cross-laminated sets are frequently capped by thin beds of green, shaly, very fine sandstone with ripples and tiny burrows. Minor structures include parallel lamination, large-scale trough cross-bedding, small-scale planar cross-lamination, soft-sedimentation deformation, and scours.

## Shale

A distinctive shale (Figure 11-2193) is present in six cores (#1107, #1108, #1109, #1126, #1128, and #1134). The shale is thin (3-15 cm, but 37 cm in #1126), green (but black in core #1108), and micaceous. Slickensides along bedding planes give the rock a waxy feeling. It also contains thin laminae of rippled very fine sandstone, producing a wavy bedding.

## Interpretations

We interpret the upper Big Injun unit to have been deposited in a braided stream with high, though widely fluctuating discharge (compare with sedimentary models of Walker and Cant, 1984; Rust and Koster, 1984; Collinson, 1986). Characteristics common to such fluvial sediments include coarse grain size, clast support with little mud, scour structures and lag deposits, numerous thin fining-upward cycles, and large-scale cross-bedding (Figures 5 and 6). Four cores display a vague overall fining-upward sequence, but the other six show no well defined vertical trend in texture.

Subfacies 4 accumulated on channel floors. Several of the thicker channel fills contain multiple scour surfaces, which suggest that they accumulated through a series of flood events. When a channel eventually choked with sediment, clays settling from suspension alternated with very fine sand migrating in the form of rip-

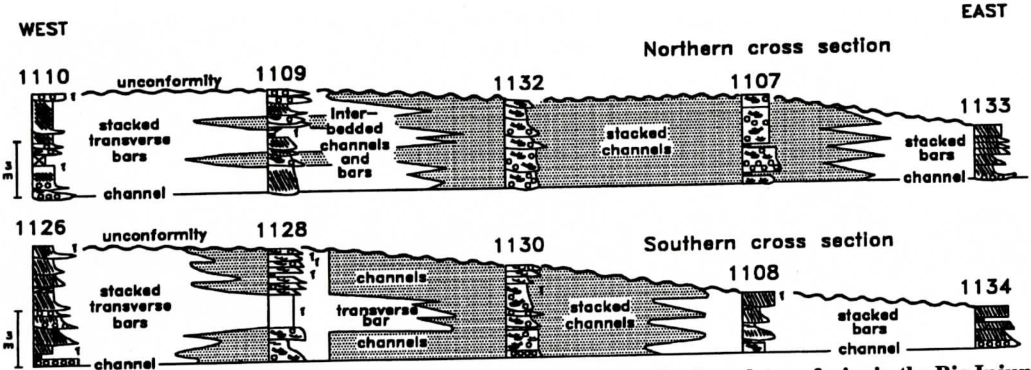


**Figure 13. Two west-to-east cross sections of the upper, coarse-grained sandstone facies in the Big Injun member. Two subfacies are recognized, channel-lag conglomerate (shaded) and transverse-bar sandstone, which stack vertically in the stratigraphic section and interfinger laterally. Lines of section are shown on Figure 2.**

ples; together these fines formed a cap with wavy bedding. Subfacies 5 represents transverse bars in which lateral migration of foreset beds generated large-scale planar cross-stratification. Trough cross-bedding and smaller sets of planar cross-bedding suggest that megaripples migrated over the bars' surface. When the bars stood emergent above surrounding channels, a wavy-bedded cap of burrowed shale partings and ripple-laminated very fine sandstone was laid down (compare with Walker and Cant, 1984).

The green, waxy shale, similar to the shaly caps of the channel-floor and transverse-bar deposits, are thought to have accumulated on flood plains or sand flats. During major floods, when the braided stream overflowed its channels, deposition of fines occurred by vertical accretion on the adjacent flood plains, temporary islands, terraces, sand flats, and bar crests. Clasts of this shale were later reworked into the overlying sandstone.

Cross sections of the upper Big Injun unit reveal an interesting areal distribution of subfacies (Figure 13). In three eastern wells (#1108, #1133, and #1134), the section is dominated by medium sandstone with large-scale planar cross-bedding, the deposits of transverse bars. These transverse-bar sandstones, stacked atop one another, are thought to represent the sand flats of Cant (1978), formed by the coalescence of in-channel bars. In a similar fashion, the upper Big Injun of western wells #1110 and #1126 consists of amalgamated transverse-bar sandstone, a second sand flat or compound bar. By way of contrast, central wells #1107, #1130, and #1132 feature a stacking of coarse sandstone with trough cross-bedding--the channel deposits. Though certainly not stabilized, it appears that the braided channel occupied this central area for a disproportionate amount of time. The last two wells, #1109 and #1128, exhibit an interbedding of channel and bar sandstones, a lateral transition zone between these two subfacies.

The reservoir sandstone at Blue Creek oil field in Kanawha County (Figure 1) is thought to be part of a separate deltaic lobe, produced during a later constructional phase of the Big Injun system. Braided-stream facies, though, are absent at Blue Creek. Whereas the lack of any delta-plain facies in the one core is attributed to the post-Price unconformity, the preserved distributary-mouth-bar sandstones of the delta front (mouth-bar front and crest) are consistently fine and very fine grained. In contrast to Granny Creek and Rouzer, there are no laminae or thin beds of quartz pebbles, nor are there coarse back-bar sandstones. Located approximately 20 km to the west, that is, seaward in Mississippian time, this site was situated farther from the eastern source area and received no coarse sediment. Seaward progradation of the braid delta recognized in Clay County thus halted somewhere to the east of Blue Creek.

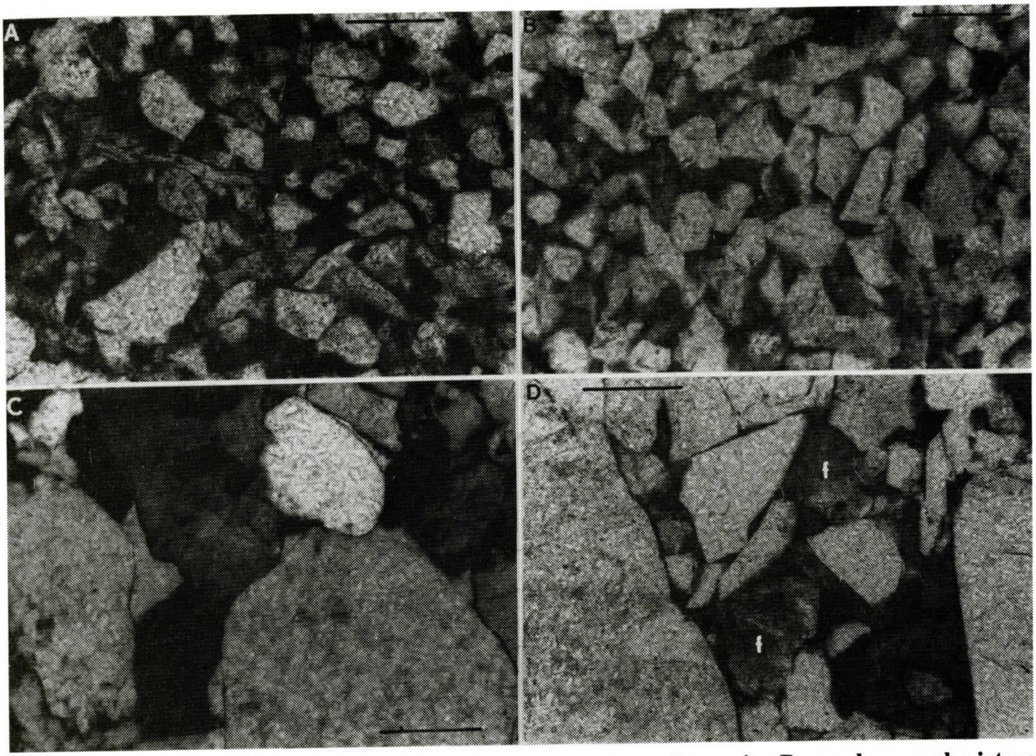


**Figure 14. Photomicrographs illustrating Big Injun lithologies and porosity. Bar scale on each picture equals 0.250 mm. *A.* Litharenite dominated by quartz (white), rock fragments of schist and shale (dark), and muscovite (large flake). Clay 1128 core, depth 2016 feet, bar-crest subfacies. *B.* Sublitharenite with abundant intergranular pore space (dark) between angular quartz. $\phi = 20.4\%$, k = 4.2 md. Clay 1109 core, depth 1908, bar-crest subfacies. *C.* Rounded, moderately sorted quartz grains are cemented by quartz overgrowths. Clay 1108 core, depth 2069 feet, transverse-bar subfacies. *D.* Primary intergranular porosity (dark) and secondary porosity within two leached plagioclase feldspars (f). $\phi = 11.7\%$, k = 19 md. Clay 1109 core, depth 1881 feet, channel-floor subfacies.**

## POROSITY AND PERMEABILITY

Petrographic analyses have been conducted on Big Injun sandstones of West Virginia by Swales (1988), Hohn and others (1993a, b), and the present authors. Quartz and chert dominate the framework grains in mouth-bar sandstones, though rock fragments of phyllite and shale are also significant (Figure 14A). Calcite, quartz, and siderite serve as the cements, but little or no clay matrix is present. Calcite also locally replaced the abundant rock fragments, creating small nodules of calcite-cemented quartz sandstone. Primary interparticle porosity is good (Figure 14B), and a well developed system of interconnected pores exists. Pore size is typically less than 0.100 mm.

Whole-core porosity and permeability data are available for seven of the 12 cores, and a plot of petrophysical data versus depth for the Clay 1109 core illustrates how these properties mirror the depositional subfacies (Figure 15). Near the base of the bar-front sandstone, porosity is 10 percent, and horizontal permeability is less than 0.1 md. Upward in this subfacies porosity increases to 16 percent, and permeability, to 0.2 md. In the overlying bar-crest sandstone, porosity continues to rise to a maximum of 22 percent, and permeability, to 24 md. The upward increase in porosity and permeability reflects the coarser grain size (larger pores), fewer shale rock fragments (which compact readily), less interstitial matrix, and fewer shale partings (which impede fluid flow). Thin zones in the

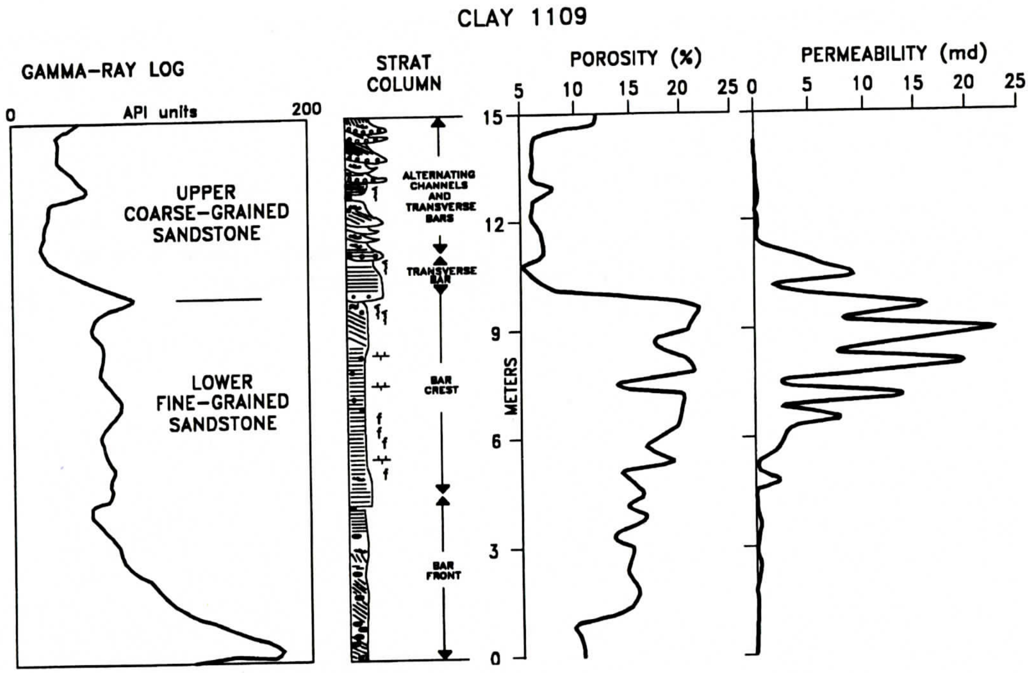


**Figure 15. Whole-core porosity and permeability values and gamma-ray log for the Big Injun sandstone in the Clay 1109 well. Note that these petrophysical properties mirror the depositional subfacies. See text for discussion.**

bar-crest sandstone with large calcite nodules have somewhat lower porosities (down to 14 percent at 7.5 meters, Figure 15) and drastically lower permeabilities (less than 9 md). Plots of petrophysical data versus depth and subfacies for the other wells of the Granny Creek field (not shown) are similar. Although absolute values are somewhat different, the stratigraphic trend is the same. Continuing upsection to the back-bar sandstone of Clay 1108 and 1126 (this subfacies is absent in Clay 1109), porosity is slightly lower than in the bar crest (16-21 percent), reflecting the slight decrease in grain size. Permeability is unpredictably variable (2-18 md) due to a complex interbedding of very fine and very coarse sandstones.

Quartz and chert constitute the major mineral components in the overlying fluvial sandstone and conglomerate (Figure 14C). Other components include rock fragments of shale, schist, and fine sandstone, as well as plagioclase and potassium feldspars. A clay matrix is locally present. Cements include calcite (in places forming small nodules), siderite, quartz, kaolinite, and pyrite. Pores are both primary interparticle and secondary molds of feldspar, and their size ranges up to 0.300-0.400 mm (Figure 14D).

Petrophysical logs for the Clay 1109 core (Figure 15) show that the basal transverse-bar sandstone has a porosity of less than 20% and a permeability of less than 10 md, considerably lower than that of the underlying mouth-bar sandstone. Although porosity is poor in this fluvial subfacies, permeability is fair because of the relatively large pores and throats. In the overlying (and alternating) bar and channel sediments, however, porosity falls to 6-12% and permeability drops drastically to 0.3 md. Porosity and permeability are reduced chiefly by calcite cement, carbonate material that probably originated in and migrated down from the overlying Greenbrier Limestone.

Overall relationships between petrophysical data and subfacies for the seven Big Injun cores are as follows. Transverse-bar sandstone demonstrates the greatest range of reservoir properties ($\phi$ = 5-19%, k = 0-36 md), owing to their extreme variability in grain size, sorting, and cements. Channel-lag conglomerate has poor porosity (mean of 7.8%), but permeability can be surprisingly good (up to 28 md) because of the large pores and connecting throats. However, fluvial sandstone and conglomerate situated immediately beneath the Greenbrier Limestone are well cemented by calcite.

Of the distributary-mouth sandstones, bar-front sediment has poor to good porosity varying from 7 to 24 percent (mean of 18.4%), but permeability is usually low (2.5 md) because of its very fine grain size. As grain size increases in the bar-crest sandstone, petrophysical variables likewise improve ($\phi$ = 10-23% with a mean of 18.9%, k $\leq$ 24 md). This subfacies is amenable to secondary oil recovery by water flooding, which requires a minimum porosity of 10% and permeability of 3 md (Watts and Overbey, 1970). Where present, calcite nodules significantly lower the reservoir's quality. Back-bar sediment, with an interbedding of fine and very coarse sandstones, exhibits an intermediate porosity (16.0%) and a good permeability (9.2 md).

## CONCLUSIONS

The coarse grain size and overall coarsening-upward texture of the Big Injun sandstone, together with its shallow-water sedimentary structures, marine trace and body fossils, yet strong fluvial influence, reflect braid-delta sedimentation. Three subfacies have been recognized in the distributary-mouth bar, collectively producing a coarsening-upward sequence on the delta front. Bar-front sandstone, deposited at and above wave base, consists of very fine sand with parallel lamination. Bar-crest sandstone accumulated at or near sea level; it is fine grained and displays small-scale trough cross-lamination. Back-bar sediment is transitional with fluvial sediments: parallel-laminated fine sand interbedded with coarse sand showing channel- fill cross-lamination.

The braided river system of the delta plain contains two subfacies arranged in distributary-channel cycles. Channel lags consist of pebbly sand with trough cross-bedding. Transverse-bar sandstone is coarse grained and exhibits large-scale planar cross-lamination. Rare shale intercalations formed when streams temporarily overflowed their channels.

In terms of reservoir quality, the bar-crest sandstone consistently records the highest porosity and permeability values. Furthermore, of the five subfacies the bar-crest sandstone is the only one economically viable for secondary oil recovery by water flooding.

## ACKNOWLEDGMENTS

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